

Geological Survey of Canada
Bulletin 382

CAMERON RIVER AND BEAULIEU RIVER
VOLCANIC BELTS OF THE ARCHEAN
YELLOWKNIFE SUPERGROUP,
DISTRICT OF MACKENZIE,
NORTHWEST TERRITORIES

M.B. Lambert

1988



Energy, Mines and
Resources Canada

Énergie, Mines et
Ressources Canada

© Minister of Supply and Services Canada 1988

Available in Canada through

authorized bookstore agents and other bookstores

or by mail from

Canadian Government Publishing Centre
Supply and Services Canada
Ottawa, Canada K1A 0S9

and from

Geological Survey of Canada offices:

601 Booth Street
Ottawa, Canada K1A 0E8

3303-33rd Street N.W.,
Calgary, Alberta T2L 2A7

A deposit copy of this publication is also available for reference
in public libraries across Canada

Cat. No. M42-382 E Canada: \$20.00
ISBN 0-660-13049-1 Other countries: \$24.00

Price subject to change without notice

Cover Photo

Deformed pillow lavas in andesite member of the Tumpline
Basalt east of Sharrie Lake.

Critical Reader

W.R.A. Baragar

Revised manuscript received — 1987-05

Final version approved for publication — 1987-11

Preface

Archean greenstone belts host gold deposits in many areas of the Canadian Shield. The Yellowknife area has been a major gold producer and an active area for exploration since the early 1930s.

This report documents the stratigraphy, structure and volcanology of polydeformed and metamorphosed Archean volcanic belts that border the eastern side of the Yellowknife supracrustal basin. The eruption processes, history of development and paleoenvironment of enormous volcanoes, that erupted mainly beneath oceans more than 2600 Ma ago, are unravelled.

Understanding the relationships of mineral occurrences to the anatomy and development of these volcanoes, as well as to the history and style of postvolcanic deformation, will be of value to further exploration in greenstone belts of the Slave Structural Province.

R.A. Price
Assistant Deputy Minister
Geological Survey of Canada

Préface

Des zones de roches vertes de l'Archéen renferment des gisements aurifères dans de nombreuses régions du Bouclier canadien. La région de Yellowknife a produit beaucoup d'or et fait l'objet de travaux d'exploration intenses depuis le début des années 30.

Le présent rapport décrit la stratigraphie, la structure et la volcanologie des zones volcaniques polydéformées et métamorphosées d'âge archéen qui bordent le côté est du bassin supracrustal de Yellowknife. Les processus d'éruption, l'évolution et le paléoenvironnement des énormes volcans qui ont fait éruption principalement dans les océans il y a plus de 2600 Ma sont mis à jour.

Comprendre les relations qui existent entre les manifestations minérales d'une part, et l'anatomie et l'évolution de ces volcans ainsi que l'évolution et le mode de la déformation post-volcanique d'autre part, facilitera les futurs efforts d'exploration dans les zones de roches vertes de la province structurale des Esclaves.

R.A. Price, sous-ministre adjoint,
Commission géologique du Canada



CONTENTS

xi	Abstract/Résumé
1	Summary/Sommaire
20	Introduction
20	Location and access
20	Physical features
20	Previous investigations
22	Present investigations
22	Geological setting
23	Geochronology
26	Acknowledgments
26	Granitic rocks
27	Sleepy Dragon Complex
29	Plutonic rocks
29	Amacher Granite
29	Detour Granodiorite
30	Defeat Plutonic Suite
31	Meander Lake Plutonic Suite
31	White muscovite-biotite granite
31	Redout Granite
32	Prosperous Granite
32	Beaulieu Group
33	Cameron River subarea
34	Cameron River Basalt
34	Definition, thickness and contact relations
34	Stratigraphy
34	Mafic lavas and related rocks
36	Lithology
37	Mafic volcanoclastic sediments
37	Shale and siltstone
38	Rhyolite fragment conglomerate and volcanoclastics
39	Interpretation
39	Webb Lake Andesite
39	Definition, distribution and thickness
39	Contact relations
40	Internal structures
41	Stratigraphy
42	Lithology
42	Felsic andesites
43	Quartz-phyric andesites
43	Mafic andesites
43	Rhyolite breccia
43	Interpretation
43	Rhyolite
44	Dome south of Fenton Lake
44	Southern dome near Dome Lake
44	Contact relations
44	Primary structures
44	Tectonic features
44	Lithology
45	Interpretation
45	Northern dome and breccias near Dome Lake
45	Dome Lake Basalt
46	Raquette Lake Formation
47	Tumpline Lake subarea
47	Tumpline Basalt
47	Definition, distribution and thickness
47	Contact relations
47	Deformation of primary structures
49	Stratigraphy
49	Turnback Lake area

51	Tumpline Lake area
51	Sharrie Lake area
52	Inclusions
52	Lithology
52	Basalts
53	Andesite member
53	Altered andesites (felsic pillows)
54	Interpretation
54	Turnback Rhyolite
54	Definition, distribution and thickness
54	Contact relations
55	Stratigraphy
55	Lavas
55	Volcaniclastic rocks
56	Lithology
57	Recognition of pyroclastic rocks
58	Tectonically deformed rhyolites
58	Interpretation
59	Sharrie Rhyolite
60	Interpretation
60	Sunset Lake subarea
60	Ultramafic rocks and biotite schist
63	Sunset Lake Basalt
63	Definition and distribution
63	Contacts
63	Primary structures
63	Layering
63	Pillows and pillow breccias
63	Massive lavas
63	Structures resulting from tectonic deformation
63	Flattened and sheared pillow lavas
64	Pseudo-layering
64	Lithology
64	Medium to dark green amphibolite
64	Pale green amphibolite
65	Very dark green amphibolite
65	Volcanic sediments
65	Conglomerate
66	Iron-formation
67	Interpretation
67	Alice Formation
67	Definition, distribution and thickness
67	Contact relations
69	Stratigraphy and lithology
69	Andesite member
70	Dacite member
72	Interpretation
72	Rhyolite
72	Contact relations
74	Rhyolite near Lake 1052
74	Dome at south end of Sunset Lake
74	Dome complex north of Sunset Lake
74	Breccias and conglomerates
75	Pyroclastic rocks
75	Dykes and sills
76	Interpretation
76	Payne Lake Formation
76	Contact relations
77	Stratigraphy and lithology
77	Mafic volcanic sediments
77	Metarhyolite
78	Felsic volcaniclastics

78	Intravolcanic sediments
78	Interpretation
79	Carbonates
79	Internal structures
79	Lithology
80	Interpretation
80	Dykes and sills
81	Amphibolite intrusions within volcanic belts
81	Distribution and thickness
81	Contact relations
83	Internal structures
83	Lithology
84	Effect of intrusions on structure and stratigraphy
85	Interpretation
85	Intrusions in granitoid gneisses
85	Cameron River subarea
85	Sunset Lake subarea
86	Boundary relationships
86	Dykes
86	Size, distribution and contact relations
86	Dyke lithology and relative chronology
87	Discussion
87	Proterozoic mafic dykes
88	Felsite dykes
88	Metamorphism
88	Regional metamorphism
88	Textures and mineralogy of metamorphosed volcanic rocks
88	Metarhyolites and metadacites
90	Meta-andesites and metabasalts
91	Amphibolite dykes
91	Metavolcaniclastic rocks, sediments and carbonates
91	Summary
92	Contact metamorphism
92	Discussion
93	Petrochemistry
93	Sampling and analysis
93	Alteration
95	Geochemical classification
96	Chemical variations within the volcanic belts
96	SiO ₂ variation diagrams
96	Major oxides
98	Minor elements
99	Change of chemistry with stratigraphy
100	Discussion
105	Structural Geology
105	Folds
109	Early folds
109	Late folds
109	Folds in the Tumpline Lake subarea
110	Folds in the Sunset Lake subarea
111	Folds and deformation in the Cameron River subarea
111	Archean faults and shear zones
112	Prevolcanic faults
112	Syn-volcanic faults
113	Shear zones and syndeformational faults
114	Sleepy Dragon shear zone
114	Amacher shear zone
115	Payne shear zone
115	Stratiform shear zones
115	Proterozoic faults
116	Summary and evolution of the Cameron-Beaulieu volcanic belts
116	Volcanism in the Cameron River subarea

116	Speculation on eruptive centres, eruption type and paleomorphology
116	Fissure eruptions
118	Central eruptions
118	Paleomorphology
120	Volcanism in the Tumpline Lake subarea
120	Eruptive centres
122	Volcanism in the Sunset Lake subarea
122	Chronicle of structural and magmatic events
123	Deformation
124	Plutonism
125	Tectonic models
126	Magma origin and emplacement
126	Timing of volcanism
127	References

Appendices

132	1. Mode and texture of granitic rocks near the Cameron River and Beaulieu River volcanic belts.
133	2. Chemical analyses, specific gravity, normative compositions, and SiO ₂ variation diagrams of rocks from the Cameron River-Beaulieu River volcanic belts.
133	A. Major oxides and specific gravity
135	B. Minor and trace elements
138	C. Normative compositions
140	D. SiO ₂ variation diagrams

Tables

24	1. Table of Formations
90	2. Summary of mineralogy and textures of metavolcanic and metasedimentary rocks of the Beaulieu Group and the Burwash Formation.
94	3. Proportions of least altered to total samples analyzed in formations from the Cameron-Beaulieu volcanic belts.

Figures

21	1. Generalized geology of the Slave Province.
(in pocket)	2. Geology of the Cameron River and Tumpline Lake subareas of the Cameron River and Beaulieu River volcanic belts.
(in pocket)	3. Geology of the Sunset Lake subarea of the Cameron River and Beaulieu River volcanic belts.
23	4. Relationship between specific gravity and SiO ₂ content, normative colour index and normative plagioclase for rocks from the Cameron and Beaulieu volcanic belts.
25	5. Radiometric dates selected mostly from units associated with the Cameron-Beaulieu volcanic belts in the southern part of the Slave Province.
27	6. Modal composition of plutonic rocks near the Cameron and Beaulieu volcanic belts.
33	7. Proportions of volcanic lithologies in the Cameron River, Tumpline Lake and Sunset Lake subareas.
(in pocket)	8. Pseudostratigraphy of volcanic belt in the Cameron River subarea.
34	9. Cameron River Basalt showing pillow lavas and pillow breccia.
35	10. Cameron River Basalt showing elongate pillows and tectonically elongated fragments in breccia.
35	11. Thin selvages and vesicle zones in pillow lava of the Cameron River Basalt.
48	12. Distribution of Turnback Rhyolite and Sharrie Rhyolite in the Tumpline Lake subarea.
49	13. Distribution of the Tumpline Basalt.
50	14. Deformed pillow lavas in andesite member of the Tumpline Basalt east of Sharrie Lake.
51	15. Columnar joints in felsic andesites of the Tumpline Basalt.
52	16. Layered units in mafic volcanoclastic rocks of the Tumpline Basalt southeast of Turnback Lake.

- 52 17. Basalt pillow lavas mould around a felsite block in the Tumpline Basalt northwest of Sharrie Lake.
- 52 18. Deformed rhyolite breccia within mafic volcanoclastics of the Tumpline Basalt north of Sharrie Lake.
- 55 19. Rounded pumice blocks in ash-flow tuffs of the Turnback Rhyolite on the southeast side of Turnback Lake.
- 58 20. Tectonically deformed rhyolite fragment breccia having pseudo eutaxitic foliation in Turnback Rhyolite, north of Goose Lake.
- 61 21. Distribution of Sunset Lake Basalt and Payne Lake Formation; reference and chemical analysis localities.
- 62 22. Boundary zone between Beaulieu River volcanic belt and Sleepy Dragon Complex near Amacher Lake.
- 66 23. Polymictic conglomerate in Sunset Lake Basalt west of Amacher Lake.
- 66 24. Deformed iron-formation near Amacher Lake.
- 67 25. Photomicrograph showing chert-magnetite iron-formation near Amacher Lake.
- 68 26. Distribution of the Alice Formation; reference and chemical analysis localities.
- 69 27. Microlitic textures in andesite lavas of the Alice Formation.
- 70 28. Lenticular protomylonite in Alice Formation andesite west of Sunset Lake.
- 70 29. Seriate feldspar-phyrlic felsic andesite of the Alice Formation.
- 71 30. Sheared rocks from the dacite member of the Alice Formation, northwest side of Sunset Lake.
- 73 31. Index map showing distribution of rhyolites in Sunset Lake subarea, reference and chemical analysis localities, and palinspastic reconstruction of rhyolite complex north of Sunset Lake.
- 75 32. Rhyolite boulder conglomerate in rhyolite complex north of Sunset Lake.
- 75 33. Rhyolite lapilli tuff.
- 79 34. Textures and structures in carbonate unit, north of Turnback Lake.
- 82 35. Amphibolite dyke swarms in Cameron River and Sunset Lake subareas.
- 84 36. Highly strained amphibolite containing inclusions of felsite.
- 88 37. Fresh quartz-bearing pigeonite diabase from post-Archean dyke cutting the Sunset Lake belt south of Payne Lake.
- 89 38. Distribution of granitic plutons, basement and metamorphic isograds near the Cameron and Beaulieu volcanic belts.
- 94 39. Least altered and altered rocks from the Cameron-Beaulieu volcanic belts plotted on Ab-Or-An diagram.
- 95 40. Volcanic formations of the Cameron-Beaulieu volcanic belts plotted on (a) AFM diagram, (b) Jensen cation plot and (c) Al_2O_3 -normative plagioclase diagram.
- 96 41. Composition of volcanic rocks of the Cameron and Beaulieu volcanic belts.
- 97 42. Subalkaline character of volcanic rocks of the Cameron and Beaulieu volcanic belts.
- 98 43. SiO_2 variation diagrams for major oxides (wt. %) in least altered volcanic rocks of the Cameron and Beaulieu volcanic belts.
- 99 44. SiO_2 variation diagrams for minor elements (wt. %) in least altered volcanic rocks of the Cameron and Beaulieu volcanic belts.
- 100 45. SiO_2 variation diagrams for minor elements (ppm) in least altered volcanic rocks from the Cameron and Beaulieu volcanic belts.
- 102 46. Major oxides and minor elements plotted against columnar sections 5, 7 and 8 of the Cameron River subarea.
- 104 47. Major oxides and minor elements in volcanic rocks from the Tumpline Lake subarea.
- 106 48. Major oxides and minor elements in volcanic rocks from the Sunset Lake subarea.
- 108 49. Classification of least altered basalts from the Cameron and Beaulieu volcanic belts according to possible tectonic environment.
- 112 50. Models of deformation of pillow lavas by processes of pure and simple shear.
- 113 51. Faults, shear zones and major dyke swarms in the Cameron - Beaulieu volcanic belts.
- 117 52. Volcanism in the Cameron River subarea.
- 119 53. Volcanism in the Tumpline Lake subarea.
- 121 54. Volcanism in the Sunset Lake subarea.
- 123 55. Generalized chronology of events near the Cameron and Beaulieu volcanic belts.
- 123 56. Polydeformed volcanic belt in the Tumpline Lake Subarea.



CAMERON RIVER AND BEAULIEU RIVER VOLCANIC BELTS OF THE ARCHEAN YELLOWKNIFE SUPERGROUP, DISTRICT OF MACKENZIE, N.W.T.

Abstract

The Cameron River and Beaulieu River volcanic belts represent Archean volcanoes deformed around granitoid basement of the Sleepy Dragon Complex. They comprise dominantly (70-85 %) tholeiitic basalt pillow lavas and breccias and lesser amounts of calc-alkaline andesites, felsic lavas, domes and tuffs. Local strongly bimodal successions contain almost equal amounts of basalt and rhyolite. Amphibolite dykes and sills form profuse swarms in both the volcanic and adjacent granitoid basement. They reflect extensive fracture systems in the sialic crust, were conduits feeding enormous volumes of surface lavas, and form an integral part of the stratigraphy in the Cameron River belt.

Compositions of volcanic units evolved with time from mafic tholeiites to felsic calc-alkaline rocks. Eruptions commenced with voluminous fissure eruptions of basaltic pillow lavas. Without major breaks in volcanic activity, eruptive style changed to central eruptions that locally built emergent piles and culminated with explosive eruption of ash-flow tuffs. The waning stages of volcanism saw the rise of rhyolite domes and some eruptions of basalt. The volcanic belts could have developed in less than one million years.

Structural and magmatic events where the Cameron-Beaulieu volcanic belts developed, involved (1) extension of submerged sialic crust, emplacement of mafic dykes, volcanism, and diachronous sedimentation; (2) deformation of supracrustals by horizontal shortening and thrusting; (3) metamorphism that peaked at amphibolite grade during major fold deformation; (4) granitic plutonism from the late stages of volcanism, through post-deformation by folding, and (?) uplift of granitic basement into deforming supracrustals; (5) mild regional shortening and folding; (6) emplacement of Proterozoic dykes; and (7) wrench faulting that postdated all the previous volcanic and structural events.

Résumé

Les zones volcaniques de Cameron River et Beaulieu River représentent des volcans archéens déformés autour du socle granitoïde du complexe de Sleepy Dragon. Elles comprennent surtout (70-85 %) des laves en coussins et des brèches de basalte tholéitique, et en moindre quantité, des andésites calco-alcalines, des laves felsiques, des dômes et des tufs. Des successions locales fortement bimodales contiennent des quantités presque égales de basalte et de rhyolite. Des dykes et des filons-couches d'amphibolite forment d'abondants essaims à la fois dans les roches volcaniques et le socle adjacent granitoïde. Ils reflètent de vastes réseaux de fractures dans la croûte sialique, qui ont servi de conduits volcaniques par lesquels sortaient d'énormes volumes de laves de surface et forment une partie intégrale de la stratigraphie de la zone de la Cameron River.

Les compositions des unités volcaniques ont évolué avec le temps, allant des roches tholéitiques mafiques aux roches calco-alcalines felsiques. Les éruptions ont commencé par de volumineuses éruptions fissurales de laves basaltiques en coussins. Sans interruptions importantes dans l'activité volcanique, le style éruptif a changé en éruptions centrales qui ont construit localement des cônes volcaniques et culminé avec une éruption explosive d'ignimbrite. Les derniers stades du volcanisme ont vu la formation de dômes de rhyolite et quelques éruptions de basalte. Les zones volcaniques ont pu se former en moins d'un million d'années.

Les événements structuraux et magmatiques pendant lesquels les zones volcaniques Cameron-Beaulieu se sont constituées, impliquaient (1) une extension de la croûte sialique submergée, une mise en place de dykes mafiques, d'un volcanisme et une sédimentation diachrone; (2) une déformation de roches supracrustales par un raccourcissement horizontal et du charriage; (3) un métamorphisme dont l'intensité maximale a atteint le degré des amphibolites pendant la principale déformation plissée; (4) du plutonisme granitique issu des dernières phases du volcanisme jusqu'à la période ultérieure de déformation sous forme de plissements et (?) un soulèvement du socle granitique en déformant des roches supracrustales; (5) un léger raccourcissement régional et plissement; (6) une mise en place de dykes protérozoïques; et (7) des failles de décrochement ultérieures à tous les événements volcaniques et structuraux antérieurs.



SUMMARY

The Cameron River and Beaulieu River volcanic belts lie in the southern part of the Slave Structural Province (centred 80 km northeast of Yellowknife, Fig. 1), which comprises highly deformed and variably metamorphosed supracrustal rocks of Archean age that occur between extensive complexes of granitic rocks. Most of the supracrustal succession belongs to the Yellowknife Supergroup (Henderson, 1970) which comprises thick sequences of volcanics, generally overlain by greywacke-mudstone turbidites derived from a mixed felsic volcanic and granitic source. The ratio of sediments to volcanics is about 4:1. Some twenty "greenstone" belts (volcanic belts metamorphosed to greenschist and amphibolite grade), with northerly to northeasterly trends lie between sedimentary basins and the granitic complexes. There is no evidence that volcanics extend any distance into the basins beneath the sediments (Henderson, 1981; McGrath et al., 1983). Belts in the southern parts of the Slave Province are dominantly thick successions of mafic pillow lavas with minor amounts of felsic tuffs, breccias and flows. Rare ultramafic units are known in the volcanic belts at 3 localities. Volcanics have been interpreted to lie unconformably on Archean granitoid crust in only a few areas.

The Cameron-Beaulieu volcanic belts are deformed around basement terrane known as the Sleepy Dragon Complex (Henderson, 1985). Both the basement and the supracrustal rocks were intruded by extensive swarms of mafic dykes and later by a series of granitic to tonalitic plutons. The Table of Formations (Table 1) applies only to the area mapped during this study.

Geochronology. Radiometric data from areas regarded as basement to the Yellowknife Supergroup indicate ages close to 3 Ga. At Point Lake, a U-Pb concordia intercept gives an age of 3152 Ma (Krogh and Gibbins, 1978; Henderson et al., 1982). Tonalitic gneiss near the Bear-Slave boundary 300 km north-northwest of Yellowknife (Grenville Lake region) yields a Rb-Sr whole-rock isochron age of 2939 ± 51 Ma ($Sr_i = 0.700 \pm 0.001$; recalculated from 3202 Ma: Frith et al., 1977) and a zircon age of $2989 + 6/-5$ Ma (Frith et al., 1986). Tonalitic boulders from a diatreme cutting volcanic rocks at Yellowknife yield zircon ages of 3210 and 3030-3040 Ma (Nikic et al., 1980) and whole-rock Sm-Nd dates of 3340 and 3390 Ma respectively (Bibicova et al., in press). The oldest known Archean basement in the Bear-Slave Province is 3480 Ma (U-Pb age on zircon S.A. Bowering, pers. comm., 1986; Bowering and van Schmus, 1984) granitoid gneiss of the Exmouth Massif in the southwestern part of the Redrock Lake area (St-Onge et al., 1984).

SOMMAIRE

Les zones volcaniques de Beaulieu River et de Cameron River se trouvent dans la partie sud de la province structurale des Esclaves (centrée à 80 km au nord-est de Yellowknife, fig. 1); celle-ci comprend des roches supracrustales très déformées et d'un métamorphisme variable de l'Archéen se trouvant entre de vastes complexes de roches granitiques. La plupart de la succession de roches supracrustales appartient au supergroupe de Yellowknife (Henderson, 1970) qui comprend d'épaisses séquences de roches volcaniques, généralement recouvertes par des dépôts de courant de turbidité et de grauwaacke, dérivés d'une origine mixte volcano-felsique et granitique. Le rapport entre les sédiments et les roches volcaniques est d'environ 4:1. Environ vingt zones de « roches vertes » (zones volcaniques métamorphosées jusqu'au degré des schistes verts et des amphibolites), avec des directions nord à nord-est se trouvant entre des bassins sédimentaires et les complexes granitiques. Il n'y a aucune preuve que des roches volcaniques s'étendent sur une quelconque distance dans le bassin sous les sédiments (Henderson, 1981; McGrath et coll., 1983). Les zones dans les parties sud de la province des Esclaves sont surtout d'épaisses successions de laves mafiques en coussins avec de petites quantités de tufs, de brèches et de coulées felsiques. On connaît peu d'unités ultramafiques dans les zones volcaniques à ces trois endroits. Ce n'est que dans quelques secteurs que l'on a interprété que les roches volcaniques reposaient en discordance sur la croûte granitoïde de l'Archéen.

Les zones volcaniques Cameron-Beaulieu sont déformées autour d'un terrane de socle connu sous le nom de complexe de Sleepy Dragon (Henderson, 1985). Les roches du socle et supracrustales ont été toutes les deux pénétrées par d'importants essaims de dykes mafiques et plus tard par une série de plutons granitiques à tonalitiques. Le tableau des formations (tableau 1) s'applique seulement à la région cartographiée pendant cette étude.

Géochronologie. Les régions considérées comme le socle du supergroupe de Yellowknife ont des âges proches de 3 Ga selon les données radiométriques. À Point Lake, une intersection sur la courbe Concordia U-Pb donne un âge de 3152 Ma (Krogh et Gibbins, 1978; Henderson et coll., 1982). Du gneiss tonalitique près de la limite entre le Grand lac de l'Ours et des Esclaves à 300 km au nord-nord-ouest de Yellowknife (région de Grenville Lake) a, selon une datation sur la courbe isochrone par le Rb-Sr sur la roche entière, 2939 ± 51 Ma ($Sr_i = 0.700 \pm 0.001$; recalculé à partir de 3202 Ma: Frith et coll., 1977) et selon une datation sur le zircon, $2989 + 6/-5$ Ma (Frith et coll., 1986). De gros blocs tonalitiques provenant d'un diatreme traversant des roches volcaniques à Yellowknife ont, d'après une datation sur le zircon, 3210 et 3030-3040 Ma (Nikic et coll., 1980) et des datations par le Sm-Nd sur la roche entière, 3340 et 3390 Ma respectivement (Bibicova et coll., sous presse). Le plus vieux socle connu archéen dans la province du Grand lac de l'Ours et des Esclaves est une gneiss granitoïde de 3480 Ma (datation par les méthodes U-Pb sur le zircon, S.A. Bowering, communication personnelle, 1986; Bowering et van Schmus, 1984) du massif d'Exmouth dans la partie sur-ouest de la région de Redrock Lake (St-Onge et coll., 1984).

Basement in the Sleepy Dragon Complex on the eastern side of the Yellowknife "basin", however, is about 120 Ma older than surrounding volcanic belts (U-Pb zircon ages of $2663 \pm 7 - 5$ Ma from the Turnback Rhyolite and $2819 \pm 40 - 31$ Ma from granitoid gneisses from Sleepy Dragon Lake; Henderson et al., 1987).

Lead isotope data from massive sulphide deposits in Yellowknife, Cameron-Beaulieu, Back River, and Hackett River volcanic belts define a paleo-isochron suggesting base metal mineralization took place at about 2670 Ma (Thorpe, 1982).

Archean granites intrusive into the Yellowknife Supergroup in the Yellowknife basin, in general range from 2.5 to 2.67 Ga (Green and Baadsgaard, 1971; Frith et al., 1977; Henderson and van Breemen, pers. comm., 1986). These plutons bracket ages of volcanism, sedimentation, deformation and metamorphism in the Yellowknife-Hearne Lake areas into the time interval between 2667-2620 Ma (Fig. 5, 55).

Granitic rocks. The Sleepy Dragon Complex forms a rectangular area between Cameron and Beaulieu rivers, comprising a metamorphosed and deformed (prior to deposition of the Yellowknife Supergroup) assemblage of mixed gneisses of diorite, tonalite and granodiorite composition that are not formally subdivided. A variety of unmetamorphosed granodiorite to granite plutons (Davidson, 1972; Henderson, 1976) that intrude the gneisses, including the Morose, Redout and Prosperous granites, are not considered part of the complex.

At one locality east of Upper Ross Lake, Henderson (1985) interpreted a schist containing granitoid pebbles, that lies between volcanic breccia and granite, as an aluminous regolith derived from weathering of the adjacent granite with which it makes sharp contact.

Although in several places contacts between volcanic and gneissic terrane are clearly exposed, they are intensely deformed in highly strained, broad, structurally complex boundary zones. Thus, possible unconformable relationships between the volcanics and gneisses have not been unequivocally documented.

Plutonic rocks within and adjacent to the volcanic belts are divided into six units whose names follow the nomenclature of Henderson (1985). Plutons are quartz-rich rocks comprising granite and granodiorite (classified according to Streckeisen, 1967). Figure 6 shows the distribution and modal compositions of the various units. Appendix 1 gives modal analyses along with various textural and mineralogical data.

Intrusions were emplaced during, or after, the main episodes of regional deformation. Some plutons

Le socle du complexe de Sleepy Dragon du côté oriental du « bassin » de Yellowknife a toutefois environ 120 Ma de plus que les zones volcaniques qui l'entourent (datation par U-Pb sur du zircon de $2663 \pm 7 - 5$ Ma provenant de la rhyolite de Turnback et de $2819 \pm 40 - 31$ Ma sur du zircon provenant de gneiss granitoïdes de Sleepy Dragon Lake; Henderson et coll., 1987).

La datation par les isotopes du plomb provenant de gisements de sulfures massifs, dans les zones volcaniques de Yellowknife, de Cameron-Beaulieu, de Back River et de Hackett River définit une paléoisochrone d'après laquelle une minéralisation en métaux communs aurait eu lieu il y a environ 2670 Ma (Thorpe, 1982).

L'âge des granites archéens intrusifs dans le supergroupe de Yellowknife à l'intérieur du bassin de Yellowknife varie, en général, de 2,5 à 2,67 Ga (Green et Baadsgaard, 1971; Frith et coll., 1977; Henderson et van Breemen, comm. pers., 1986). L'âge du volcanisme, de la sédimentation, de la déformation et du métamorphisme dans les régions de Yellowknife-lac Hearne se situe dans l'intervalle de 2667 - 2620 Ma (fig. 5, 55).

Roches granitiques. Le complexe de Sleepy Dragon forme une aire rectangulaire entre Cameron River et Beaulieu River, comprenant un assemblage métamorphisé et déformé (avant le dépôt du supergroupe de Yellowknife) de gneiss mélangés formés à partir de diorite, tonalite et granodiorite qui ne sont pas formellement subdivisés. Des plutons, dont la composition varie d'une variété de granodiorite non métamorphisée à du granite (Davidson, 1972; Henderson, 1976), qui pénètrent les gneiss, incluant les granites de Morose, de Redout et de Prosperous, ne sont pas considérés comme faisant partie du complexe.

À un endroit situé à l'amont et à l'est du lac Ross, Henderson (1985) a interprété un schiste contenant des galets granitoïdes et qui se trouve entre des brèches volcaniques et du granite, comme étant un régolite alumineux dérivé de l'altération du granite adjacent avec lequel il a un contact net.

Bien qu'à plusieurs endroits les contacts entre les roches volcaniques et le terrane gneissique soient clairement évidents, ils sont intensément déformés dans des zones limites structurellement complexes, vastes et très déformées. Ainsi la relation de discordance possible entre les roches volcaniques et les gneiss n'a pas fait l'objet d'études suffisantes.

Les roches plutoniques dans les zones volcaniques et voisines de celles-ci sont divisées en six unités dont les noms correspondent à la nomenclature de Henderson (1985). Les plutons sont des roches riches en quartz comprenant du granite et de la granodiorite (classés selon Streckeisen, 1967). La figure 6 montre la distribution et les compositions modales des diverses unités. L'annexe 1 donne les analyses modales ainsi que diverses données minéralogiques et texturales.

Les intrusions ont pris place durant, ou après, les épisodes majeurs de la déformation régionale. Certains plutons

have drastically effected the pattern of deformation in the sediments and volcanic rocks that surround them, whereas others cut across all structures. Henderson (1985) considered the Detour Granodiorite and Amacher Granite to be older than the other plutons by virtue of being metamorphosed.

The Beaulieu Group. The Beaulieu Group contains most of the volcanic belts in the southern parts of the Slave Province (Henderson, 1970, 1985). In the areas mapped, the volcanic belts are divided into three subareas, namely, (1) Cameron River subarea, (2) Tumpline Lake subarea and (3) Sunset Lake subarea. Each subarea has a unique stratigraphy and style of deformation. Legends of Figures 2 and 3 show schematic facies relations (notwithstanding deformation) in the three subareas.

Cameron River subarea. The Cameron River subarea contains 40 km of a northerly trending volcanic belt that marks the boundary between sedimentary rocks of the Burwash Formation to the west and granitic rocks of the Sleepy Dragon Complex to the east. Width of the volcanic belt varies from 3000 to 4200 m between Fenton and Webb lakes, tapers abruptly to about 50 m north of Upper Ross Lake, and pinches out near Victory Lake. Stratigraphic thickness of the belt is difficult to establish because, (1) the belt is highly deformed and present data cannot resolve the degree to which tectonic flattening, stretching and possibly fault imbrication have altered the original thickness, (2) the belt may represent a series of flows that overlap progressively westward and possibly never were a thick vertical pile, and (3) the "pile" was expanded by intrusion of a huge swarm of dykes and sills.

The eastern side of the belt is irregular where it makes faulted contacts against granitic rocks of the Sleepy Dragon Complex. The present horizontal relief of this contact is about 2000 m. In contrast, the western side of the belt is chiefly smooth and gently undulating.

The volcanic belt generally is a westward-facing homoclinal succession of metavolcanic rocks comprising four formations (Fig. 2, 8): (1) Cameron River Basalt — dominantly pillow lavas of basaltic composition, 85 % of the belt; (2) rhyolite to dacite domes, flows, breccias and arenites, 3 % of the belt; (3) Webb Lake Andesite — pillow lavas, breccias and volcanoclastic rocks, 10 % of the belt; and (4) Dome Lake Basalt — lavas and volcanoclastic rocks, 2 % of the belt.

The Raquette Lake Formation is an epiclastic unit that lies between the volcanic belt or the Sleepy Dragon Complex and the Burwash sediments.

ont affecté sérieusement le modèle de la déformation dans les roches sédimentaires et volcaniques situées à la périphérie, tandis que d'autres affectant toutes les structures. Henderson (1985) considère que la granodiorite Detour et le granite Amacher sont plus âgés que les autres plutons à cause de leur métamorphisme.

Le groupe de Beaulieu. Le groupe de Beaulieu contient la plupart des zones volcaniques de la partie sud de la province des Esclaves (Henderson, 1970, 1985). Dans les régions levées, les zones volcaniques sont divisées en trois sous-régions, (1) la sous-région de Cameron River, (2) la sous-région de Tumpline Lake et (3) la sous-région de Sunset Lake. Chacune des sous-régions a une stratigraphie propre et un style bien défini de déformation. Les légendes des figures 2 et 3 montrent schématiquement les relations des faciès dans les trois sous-régions (la déformation n'est pas représentée).

La sous-région de Cameron River. La sous-région de Cameron River contient une zone volcanique d'orientation nord de 40 km qui indique la limite entre les roches sédimentaires de la formation de Burwash à l'ouest et les roches granitiques du complexe de Sleepy Dragon à l'est. La largeur de la zone volcanique varie de 3000 à 4200 m entre les lacs Fenton et Webb, elle s'amincit brusquement à environ 50 m au nord du lac Upper Ross et disparaît près du lac Victoria. La puissance de la stratigraphie de la zone est difficile à établir car (1) la zone est très déformée et les données actuelles ne peuvent pas résoudre le degré auquel l'aplatissement tectonique, l'étirement et l'imbrication possible de failles ont altéré l'épaisseur d'origine, (2) la zone peut représenter une série de coulées qui se sont chevauchées progressivement vers l'ouest et probablement n'ont jamais été sous forme d'édifice vertical épais et (3) « l'édifice » s'est agrandi par intrusion d'un énorme essaim de dykes et de filons-couches.

Le côté oriental de la zone est irrégulier et a des failles avec les roches granitiques du complexe de Sleepy Dragon. Le relief horizontal de ce contact a environ 2000 m. Au contraire, le côté occidental de la zone est surtout uniforme et légèrement ondulée.

La zone volcanique est en général une succession homoclinal faisant face à l'ouest de roches metavolcaniques comprenant quatre formations (fig. 2, 8): (1) le basalte de Cameron River — surtout les laves en coussins de composition basaltique, représentant 85 % de la zone; (2) des dômes, des coulées et des brèches de rhyolite à dacite et des arénites, représentant 3 % de la zone; (3) l'andésite de Webb Lake — laves en coussins, brèches et roches volcanoclastiques, représentant 10 % de la zone; et (4) basalte de Dome Lake — laves et roches volcanoclastiques, représentant 2 % de la zone.

La formation de Raquette Lake est une unité épicastrale qui se trouve entre la zone volcanique ou le complexe de Sleepy Dragon et les sédiments de Burwash.

The Cameron River Basalt consists dominantly of basaltic pillow lavas, pillow breccias and minor massive lavas. East of Allan and Fenton lakes the formation contains a volcanoclastic member of basaltic breccias, pillow breccias and basaltic arenites, lentils of rhyolite boulder conglomerate, breccia and arenites, and lentils of shale. Apparent thickness is 2000-3300 m in the northern part, where the formation makes up the entire width of the belt. It thins abruptly to 700 m, 4 km north of Upper Ross Lake, then tapers gradually and pinches out south of the lake.

The formation is structurally unconformable against the Sleepy Dragon Complex, and is conformable or interfingers with the overlying Webb Lake Andesite and the Burwash Formation. Precise contact relations in this zone of interfingering are obscured by a swarm of thick mafic dykes. Large rhyolite bodies overlie the Cameron River Basalt south of Webb Lake.

The Webb Lake Andesite comprises pillow lavas and breccias of andesite and basaltic andesite and associated volcanic arenites and rudites. West and southwest of Webb Lake the formation apparently interfingers with the Cameron River Basalt across a width of 1400 m, thins to about 400 m between columns 2 and 5 (see Fig. 8) and ends about 1100 m south of column 2. The thickness is variable near the southern end where the formation overlies rhyolite domes. Along the eastern side of Allan Lake the formation pinches and swells and has a maximum thickness of 300 m.

Contacts between the Webb Lake Andesite and the Cameron River Basalt are chiefly conformable. In places the contact is marked only by an abrupt change in lithology and there is no change in the structure of the pillowed units.

Three lenticular bodies of rhyolite at or near the top of the Cameron River Basalt northeast of Dome Lake and southeast of Fenton Lake are interpreted as domes and associated flanking breccias.

The Dome Lake Basalt is a bell-shaped unit, comprising crudely stratified alternating pillowed and clastic units about 300 m thick, that protrudes from the southwestern side of the volcanic belt 3 km north-east of Dome Lake.

The Raquette Formation is a discontinuous unit comprising heterogeneous assemblages of sandstone, conglomerate and carbonate that extend for about 5 km along the western side of the volcanic belt or the Sleepy Dragon Complex between Upper Ross Lake and Raquette Lake. The formation, which has a maximum thickness of about 60 m, conformably overlies, and locally interfingers with, mafic volcanics of the Cameron River Basalt. Locally, basal conglomerates of the Raquette Formation grade into mafic breccias of the Cameron River Basalt. North of

Le basalte de Cameron River consiste surtout en laves basaltiques en coussins, brèches en coussins et en un peu de laves massives. À l'est des lacs Allan et Fenton, la formation contient un membre volcanoclastique de brèches basaltiques, de brèches en coussins et d'arénites basaltiques, des lentilles de conglomérats de galets de rhyolite, des brèches et des arénites, et des lentilles de schiste argileux. L'épaisseur apparente est de 2000 à 3300 m dans la partie nord, où la formation couvre l'entière largeur de la zone. Elle s'amincit brutalement à 700 m, à 4 km au nord et à l'amont du lac Ross, puis s'amincit graduellement et disparaît au sud du lac.

La formation est en discordance structurale par rapport au complexe de Sleepy Dragon et en concordance ou s'interdigite avec l'andésite de Webb Lake qui la recouvre et la formation de Burwash. Le type précis de contact dans cette zone d'interdigitation est caché par un essaim de dykes mafiques épais. D'importantes masses de rhyolite recouvrent le basalte de Cameron River au sud du lac Webb.

L'andésite de Webb Lake comprend des laves en coussins, des brèches d'andésite et l'andésite basaltique, et à de l'arénite et à des rudites volcaniques associées. À l'ouest et au sud-ouest du lac Webb, la formation apparemment s'interdigite avec le basalte de Cameron River sur une largeur de 1400 m, s'amincit à environ 400 m entre les colonnes 2 et 5 (voir fig. 8) et s'achève à environ 1100 m au sud de la colonne 2. L'épaisseur est variable près de l'extrémité sud où la formation recouvre des dômes de rhyolite. Le long de la rive orientale du lac Allan, la formation s'étrangle et gonfle et a une épaisseur maximale de 300 m.

Les contacts entre l'andésite de Webb Lake et le basalte de Cameron River sont surtout en concordance. À certains endroits, le contact n'est marqué que par un changement abrupt de la lithologie et il n'y a pas de changement dans la structure des unités en coussins.

Trois masses lenticulaires de rhyolite sur le basalte de Cameron River ou près de celui-ci au nord-est du lac Dome et au sud-est du lac Fenton sont interprétés comme étant des dômes et des brèches latérales associées.

Le basalte de Dome Lake est une unité en forme de cloche, comprenant des unités alternées en coussins et clastiques grossièrement stratifiées ayant environ 300 m d'épaisseur qui font saillie sur le côté sud-est de la zone volcanique à 3 km au nord-est du lac Dome.

La formation de Raquette est une unité discontinue comprenant des assemblages hétérogènes de grès, conglomérat et roches carbonatées qui s'étendent sur environ 5 km le long de la bordure occidentale de la zone volcanique ou du complexe de Sleepy Dragon entre l'amont du lac Ross et le lac Raquette. La formation qui a une épaisseur maximale d'environ 60 m, recouvre en concordance, et localement s'interdigite avec des roches volcaniques mafiques du basalte de Cameron River. Localement, des conglomérats basaux de la formation de Raquette se transforment en brèches mafiques du basalte de Cameron River. Au nord du lac

Raquette Lake the Raquette Formation is unconformable with the Sleepy Dragon Complex, but in most places the contact is faulted. Although all units are rarely present in any one area, the succession is basal conglomerate overlain by quartzite or felsic volcanic detritus, and locally overlain by calcareous mudstone. Generally, conglomerates contain abundant mafic volcanic clasts, minor felsic volcanic clasts, and rare granitoid pebbles in a carbonate matrix. In one place, where the formation makes contact with granitoid basement, mafic cobble conglomerate is overlain by granitic fragment conglomerate in which the clasts are similar to the deformed granitoids of the basement complex.

The granitic cobble conglomerates and quartzite evince erosion from the granitoid basement before and during deposition of the volcanic belt.

Tumpline Lake subarea. The strongly bimodal volcanic belt in the Tumpline Lake subarea contains almost equal amounts of rhyolite and basalt. The belt comprises three formations plus a unit of carbonate: (1) Tumpline Basalt, (2) Turnback Rhyolite and (3) Sharrie Rhyolite.

The Tumpline Basalt consists of pillowed lavas and layered volcanoclastic rocks of basaltic composition except for a minor member of andesite west of Devore Lake. The formation, which may have a maximum thickness of 1000 m, appears to lie conformably on Turnback Rhyolite and Sharrie Rhyolite and is interpreted as a wedge within the Turnback Rhyolite. The basalts are overlain conformably by metasediments of the Burwash Formation. In the easterly trending "wings" of the volcanic belt southeast of Turnback Lake, the formation is lavas and minor volcanoclastics in the western part (where the wings meet), but it is almost entirely well bedded to crudely layered volcanoclastics in the eastern end. Around the large pluton south of Tumpline Lake the formation is dominantly pillow lavas and pillow breccias in the northern part, mixed pillow lavas and layered rocks along the southern part, and dominantly layered rocks along the southwestern part. All the rocks are intensely deformed. Near Sharrie Lake the formation is almost entirely pillow lavas except near the eastern extremities of this area where it is layered and schistose amphibolite.

The Turnback Rhyolite includes all rhyolite units in this subarea that lie directly above the Tumpline Basalt near Turnback Lake and above and below the basalt south of Tumpline Lake. The Turnback Rhyolite is stratigraphically the highest formation of the Beaulieu Group in parts of this subarea. It makes up about 40% of the volcanic belt and comprises intensely deformed and metamorphosed lavas, (?) domes, breccias and volcanoclastics. Although the

Raquette, la formation de Raquette est en discordance avec le complexe de Sleepy Dragon mais dans la plupart des endroits, le contact est faillé. Bien que toutes les unités soient rarement présentes en un seul endroit, la succession est un conglomérat basal recouvert par de la quartzite ou des débris volcano-felsiques, et localement recouvert par une pélite calcareuse. En général, les conglomérats contiennent d'abondants fragments de roches volcaniques mafiques, un peu de fragments de roches volcano-felsiques, et quelques rares galets granitoïdes dans une matrice carbonatée. À un endroit, où la formation est en contact avec un socle granitoïde, un conglomérat de galets mafiques est recouvert par un conglomérat de fragments granitiques dans lesquels les fragments de roches sont semblables aux granitoïdes déformés du complexe du socle.

Les conglomérats de fragments granitiques témoignent de l'érosion du socle granitoïde avant et pendant le dépôt de la zone volcanique.

Sous-région de Tumpline Lake. La ceinture volcanique fortement bimodale dans la sous-zone de Tumpline Lake contient des quantités presque égales de rhyolite et de basalte. La ceinture comprend trois formations plus une unité de carbonate: (1) basalte de Tumpline, (2) rhyolite de Turnback et (3) rhyolite de Sharrie.

Le basalte de Tumpline consiste en laves en coussins et en roches volcanoclastiques litées de composition basaltique, sauf pour un membre peu important d'andésite à l'ouest du lac Devore. La formation qui peut avoir une épaisseur maximale de 1000 m, semble reposer en concordance sur la rhyolite de Turnback et la rhyolite de Sharrie et elle est interprétée comme un coin dans la rhyolite de Turnback. Les basaltes sont recouverts en concordance par des métasédiments de la formation de Burwash. Dans les « ailes » orientées vers l'est de la zone volcanique au sud-est du lac Turnback, la formation est formée de laves et de petites quantités de roches volcanoclastiques dans la partie occidentale (où les ailes se rencontrent), mais elle est habituellement entièrement couverte de couches régulières à grossières de roches volcanoclastiques dans l'extrémité orientale. Autour du grand pluton au sud du lac Tumpline, la formation est composée surtout de laves en coussins et de brèches en coussins dans la partie nord, de laves en coussins mélangées et de roches litées le long de la partie sud, et surtout de roches litées le long de la partie sud-ouest. Toutes les roches sont intensément déformées. Près du lac Sharrie, la formation est presque entièrement constituée de laves en coussins sauf près des extrémités orientales de cette région où elle se présente sous forme d'amphibolite litée et shisteuse.

La rhyolite de Turnback inclut toutes les unités de rhyolite de la sous-région qui reposent directement sur le basalte de Tumpline près du lac Turnback et au-dessus et en-dessous du basalte au sud du lac Tumpline. La rhyolite de Turnback est stratigraphiquement la formation la plus élevée du groupe de Beaulieu dans certaines parties de cette sous-région. Elle représente environ 40% de la zone volcanique et comprend des laves métamorphosées et intensément déformées (?), des dômes, des brèches et des roches volcanoclastiques. Bien que

maximum thickness of the formation is not known, it may be as much as 1000 m south of Tumpline Lake. The unit thins eastward from up to 250 m north of Tumpline Lake to 10-25 m in the eastern "wings" of the volcanic belt. Most of the formation contains volcanoclastics including ash-flow tuffs, bedded tuffs, breccias and arenites. Massive volcanoclastics, including non- and partly-welded ash-flow tuffs, occur where the felsic volcanic succession is thickest. In the "wings" of the belt, bedded rhyolites are interlayered with mafic pillow lavas, massive breccias and crudely bedded volcanic arenites and breccias. Some well bedded rocks are volcarenites and lutites that may be water-laid tuffs or epiclastic rocks.

The Sharrie Rhyolite comprises sparsely porphyritic lavas, (?) domes, minor pyroclastic rocks and crystal tuff. Maximum thickness is in the order of 850 m. The formation appears to lie conformably between the Tumpline Basalt and the Burwash Formation.

Sunset Lake subarea. The volcanic belt in the Sunset Lake subarea comprises three formal formations and two informal units: (1) ultramafic rocks (informal), (2) Sunset Lake Basalt, (3) Alice Formation, divided into an andesite member and a dacite member, (4) rhyolite (informal) and (5) Payne Lake Formation. Figure 3 shows facies relations in this subarea.

Ultramafic rocks are known in two localities in the Cameron River (T.M. Kusky, pers. comm., 1986) and Sunset Lake subareas. They were mapped for 3 km along the highly complex boundary zone between the Sunset Lake Basalt and granitoid gneisses of the Sleepy Dragon Complex. The unit pinches and swells, has a maximum width of 60 m and tapers out southwards into biotite schist. Contacts with adjacent dark green schistose mafic volcanics, mafic pillow lavas, amphibolite dykes and granitoid gneisses are ubiquitously strained and have a strong foliation. The ultramafite is largely a highly deformed serpentinite that shows a wide range of mineralogical and textural varieties including aphanitic serpentinite (mainly antigorite) with veins of chrysotile, fibrous actinolite, talc and chlorite schists, and relict coarse grained rocks. Some rocks contain olivine, enstatite and amphibole. The long narrow outcrop pattern, with an orientation similar to amphibolite intrusions in the boundary zone, suggests that this unit could be a dyke or sill.

The Sunset Lake Basalt which makes up about 70% of this volcanic belt chiefly comprises pillow lavas, pillow breccias and hyaloclastites, but contains minor iron-formation, conglomerate and volcanic sediments. The formation, of unknown thickness, lies below or interfingers with the Alice Formation, and with rhyolites and volcanoclastic sediments of the Payne Formation. The boundary zone between the

l'épaisseur maximale de la formation soit inconnue, elle pourrait atteindre 1000 m au sud du lac Tumpline. L'unité s'amin- cuit vers l'est, en allant de 250 m au nord du lac Tumpline à 10 à 25 m dans les « ailes » orientales de la zone volcani- que. La plupart de la formation contient des roches volcano- clastiques incluant de l'ignimbrite, des tufs lités, des brèches et de l'arénite. Des roches volcanoclastiques massives incluant de l'ignimbrite soudée en partie ou non soudée, se trouvent là où la succession volcano-felsique est la plus épaisse. Dans les « ailes » de la zone, des rhyolites stratifiés sont intercalés avec des laves mafiques en coussins, des brèches massives et des arénites volcaniques grossièrement litées et des brèches. Quelques roches bien litées sont des volcarenites et des lutites qui peuvent être des tufs déposés dans l'eau ou des roches épicalstiques.

La rhyolite de Sharrie comprend des laves occasionnel- lement porphyritiques, (?), des dômes, et de petites roches pyroclastiques et du tuf à cristaux. L'épaisseur maximale est de l'ordre de 850 m. La formation semble reposer en concordance sous le basalte de Tumpline et la formation de Burwash.

Sous-région de Sunset Lake. La zone volcanique dans la sous- région de Sunset Lake comprend trois formations régulières et deux unités irrégulières: (1) roches ultramafiques (irrégulières), (2) basalte de Sunset Lake, (3) formation d'Alice, divisée en membre andésitique et membre dacitique, (4) rhyo- lite (irrégulière) et (5) formation de Payne Lake. La figure 3 montre les relations des faciès dans la sous-région.

Des roches ultramafiques sont connues dans les deux endroits dans les sous-régions de Cameron River (T.M. Kusky, comm. pers., 1986) et de Sunset Lake. Elles ont été cartographiées sur 3 km le long de la zone limite très complexe entre le basalte de Sunset Lake et les gneiss grani- toïdes du complexe de Sleepy Dragon. L'unité se rétrécit puis gonfle, a une largeur maximale de 60 m et s'amincit pour disparaître vers le sud dans un schiste à biotite. Les contacts avec les roches volcaniques mafiques schisteuses vert foncé et adjacentes, avec des laves mafiques en coussins, des dykes d'amphibolite et des gneiss granitoïdes sont partout déformés et ont une forte schistosité. La roche ultrabasique est surtout une serpentinite très déformée que présente une vaste gamme de variétés minéralogiques et texturales incluant de la serpentinite aphanitique (surtout de l'antigorite) avec des veines de chrysotile, d'actinolite fibreuse, de talc et de chlorito-schiste, et des roches à grains grossiers résiduels. Certaines roches contiennent de l'olivine, de l'enstatite et de l'amphibole. La forme longue et étroite de l'affleurement, avec une orientation semblable à celle des intrusions d'amphi- bolite dans la zone limite, suggère que cette unité pourrait être un dyke ou un filon-couche.

Le basalte de Sunset Lake, qui représente environ 70 % de cette zone volcanique comprend principalement des laves en coussins, des brèches en coussins et des hyaloclastites, mais contient une petite formation ferrifère, des conglomérats et des sédiments volcaniques. La formation, d'une épais- seur inconnue, se trouve sous la formation d'Alice ou l'interpénètre ainsi que les rhyolites et les sédiments volca- noclastiques de la formation de Payne. La zone limite entre

Sleepy Dragon Complex and the Sunset Lake Basalt along the western side of the Beaulieu volcanic belt, is a linear to gently undulating zone of high ductile strain complicated by large amphibolite dykes and the Amacher shear zone. Fault-bound lenses of granitoid gneiss occur in highly strained Sunset Basalt within 1 km of the main volcanic contact.

The polymictic conglomerate near the base of the Sunset Basalt contains clasts of metabasalt, metadiabase, vein quartz and granitic gneiss, suggesting derivation from both the volcanic belt and granitoid terrane.

The Alice Formation is divided into two members: a lower andesite member of pillow lavas, breccias, massive flows and minor pyroclastics and volcanic sediments; and an upper dacite member of andesite to dacite tuffs, volcanic breccias, lavas and cataclastic equivalents. In most places the formation conformably overlies, or interfingers with, the Sunset Lake Basalt. At the south end of Sunset Lake a local unconformable relationship may exist between the Alice and Burwash formations. The Alice may have a maximum thickness of about 1200 m. The andesite member may be up to 800 m thick.

This formation records a major change in the composition of volcanic products, type of eruption and environment of eruption within the Sunset Lake volcanic belt. Volcanism began with quiet submarine effusion of andesite and felsic andesite pillow lavas, which gave way to massive, vesiculated lavas and explosive eruption of andesitic and dacitic tuff when the volcanic pile had built up near to, or possibly locally emergent above, sea level. Rhyolite to dacite domes may mark vents of some of the pyroclastic eruptions that formed the dacite member.

Rhyolite bodies in the Sunset Lake subarea lie at, or near the top of, the volcanic succession and commonly form a unit between the Sunset Lake Basalt and the Burwash Formation (Fig. 31). They are about 1100 m thick and occur in three main areas: (1) a thick unit near Lake 1052 that correlates with the Turnback Rhyolite, (2) a dome at the south end of Sunset Lake, and (3) a dome complex near the north end of Sunset Lake. The dome complex north of Sunset Lake comprises 3 rhyolite units superimposed on one another and locally separated by units of andesite and dacite. Aprons of pyroclastic and coarse epiclastic debris extend 2-6 km north of the complex. Rhyolite dykes and dyke swarms (2-40 m thick) in the underlying mafic and intermediate volcanics are considered to be consanguineous with the main rhyolite bodies.

le complexe de Sleepy Dragon et le basalte de Sunset Lake le long du côté occidental de la zone volcanique de Beaulieu, est une zone linéaire légèrement ondulée de terrain qui a subi une forte déformation ductile rendue complexe par la présence d'importants dykes d'amphibolite et la zone de cisaillement d'Amacher. Des lentilles de gneiss granitoïde limitées par des failles se trouvent dans les basaltes de Sunset très déformés dans le kilomètre suivant le principal contact volcanique.

Le conglomérat polygénique près de la base du basalte de Sunset contient des fragments de metabasalte, de métabasalte, une veine de quartz et de gneiss granitique, ce qui suggère qu'il dérive à la fois de la zone volcanique et d'un terrane granitoïde.

La formation d'Alice est divisée en deux membres: un membre inférieur d'andésite avec des laves en coussins, des brèches, des coulées massives et un peu de roches pyroclastiques et de sédiments volcaniques; et un membre supérieur dacitique allant des andésites aux tufs dacitiques, de brèches volcaniques, et de laves et d'équivalents cataclastiques. Dans la plupart des endroits, la formation recouvre en concordance ou interpénètre le basalte de Sunset Lake. À l'extrémité sud de Sunset Lake, une relation locale en discordance existe peut-être entre les formations d'Alice et de Burwash. La formation d'Alice peut avoir une épaisseur maximale d'environ 1200 m. Le membre d'andésite peut atteindre 800 m d'épaisseur.

Cette formation enregistre un changement majeur dans la composition de produits volcaniques, type d'éruption et environnement de l'éruption à l'intérieur de la zone volcanique de Sunset Lake. Le volcanisme a commencé par une effusion sous-marine tranquille d'andésite et de laves en coussins d'andésite felsique; peu à peu les laves sont devenues massives et vacuolaires, puis il y a eu éruption explosive de tuf andésitique et dacitique lorsque le cône volcanique se formait près du niveau de la mer, ou probablement en émergeait localement. Des dômes de rhyolite à dacite peuvent être à l'emplacement des cheminées volcaniques de quelques-unes des éruptions pyroclastiques qui ont formé le membre dacitique.

Des masses de rhyolites dans la sous-région de Sunset Lake reposent sur le sommet ou près de celui-ci de la succession volcanique et forment communément une unité entre le basalte de Sunset Lake et la formation de Burwash (fig. 31). Elles ont environ 1100 m d'épaisseur et se trouvent dans trois zones principales: (1) une unité épaisse près du lac 1052 qui est mis en corrélation avec la rhyolite de Turnback, (2) un dôme à l'extrémité sud du lac Sunset, et (3) un complexe en forme de dôme près de l'extrémité nord du lac Sunset. Le complexe en forme de dôme au nord du lac Sunset comprend trois unités rhyolitiques surimposées les unes aux autres et localement séparées par des unités d'andésite et de dacite. Des plaines d'épandage de débris pyroclastiques et épilastiques grossiers s'étendent côté nord sur 2 à 6 km du complexe. On considère que les dykes de rhyolite et les essaims de dykes (2-40 m d'épaisseur) dans les roches volcaniques mafiques et neutres sous-jacentes et les masses principales de rhyolite seraient consanguins.

The thick local accumulations of rhyolite represent eruptive centres from which the last gasp of volcanism took place in the Sunset Lake subarea. Domes developed over a period of time during which volcanic products of intermediate and felsic compositions overlapped and became interfingered.

Subaqueous environments are indicated by pillow lavas of intermediate composition that interfinger with parts of the rhyolite members. However, the proximity of the domes south of Sunset Lake with welded tuffs (Alice dacites) suggests that the top of the dome may have emerged above sea level. A subaerial or shallow water environment is also indicated by rhyolite conglomerate and dominance of tuffs amongst flanking rhyolite deposits and the overlying felsic andesites. Emergence, if it occurred, was temporary, for the rhyolite members were covered with apparent conformity with greywacke and siltstone successions of the Burwash Formation.

The Payne Lake Formation forms most of the southeastern arm of the volcanic belt in the Sunset Lake subarea. It comprises a succession of volcanoclastic sediments that are dominantly layered schists and gneisses. The succession is divided into 3 main lithologies: (1) quartz-feldspar schists with minor mica (meta-rhyolite or detritus derived from rhyolite); (2) biotite-muscovite-quartz-feldspar schists (sediments derived from felsic volcanics); (3) amphibole-rich schists with variable amounts of mica (sediments derived from mafic volcanics).

The Payne Lake Formation is interpreted mainly as epiclastic volcanic sediments derived directly from the adjacent volcanic pile. Some of the rhyolitic units, however, may be subaqueous flows or domes from which coarse rhyolitic blocks formed proximally and fine clastic sediments formed distally.

Carbonate units do not form discrete beds (sedimentary layers) in the volcano-sedimentary succession. Instead, they generally are carbonate-rich zones and lentils (up to 85 m thick) in clastic volcanic rocks (that may be primary breccias or tectonic breccias) mainly at the top of rhyolite units where they are overlain by the Burwash Formation. Carbonate forms the matrix of rhyolite breccias and conglomerates associated with rhyolite domes and flows north of Sunset Lake and of rhyolite boulder conglomerates and breccia along the volcano-sedimentary contact northeast of Ross Lake. Main units occur between Turnback and Rex lakes, near Lake 1052, and south and west of Detour Lake. Carbonate occurs as lentils within rhyolite units near Lake 1052 and at the contact between rhyolite and the Alice andesites between Sunset Lake and Lake 1052. The carbonate was emplaced during the late stages of volcanism or after volcanism.

Les épaisses accumulations locales de rhyolite représentent des centres d'éruption à partir desquels le dernier sursaut de volcanisme a eu lieu dans la sous-région de Sunset Lake. Les dômes se sont érigés pendant la même période que celle où les produits volcaniques de composition neutre et felsique se sont chevauchés et se sont interpénétrés.

Des environnements subaqueux sont indiqués par les laves en coussins de composition neutre interpénétrées par des parties des membres rhyolitiques. Toutefois, la proximité des dômes au sud du lac Sunset avec des ignimbrites (dacites d'Alice) suggère que le haut du dôme peut avoir émergé de la mer. Un environnement d'eau peu profonde ou subaérien est aussi indiqué par un conglomérat de rhyolite et une prédominance de tufs parmi les dépôts latéraux de rhyolite et les andésites felsiques qui les recouvre. L'émergence, si elle s'est produite, a été temporaire car les membres rhyolitiques étaient couverts apparemment en concordance, de succession de grauwaque et de siltstone de la formation de Burwash.

La formation de Payne Lake forme la plupart du bras sud-est de la zone volcanique dans la sous-région de Sunset Lake. Elle comprend une succession de sédiments volcanoclastiques qui sont surtout des gneiss et des schistes lités. La succession est divisée en trois lithologies principales: (1) des schistes quartzo-feldspatiques avec peu de mica (métarhyolite ou débris dérivés de rhyolite); (2) des schistes quartzo-feldspatiques à biotite et à muscovite (sédiments dérivés de roches volcaniques felsiques); (3) des schistes riches en amphibole avec des quantités variables de mica (sédiments dérivés de roches volcaniques mafiques).

La formation de Payne Lake est interprétée principalement comme étant des sédiments volcaniques épicaustiques dérivés directement de l'édifice volcanique adjacent. Certaines unités rhyolitiques toutefois, peuvent être des coulées subaqueuses ou des dômes à partir desquels des blocs rhyolitiques grossiers se sont formés proximalelement et des sédiments clastiques fins formés distalelement.

Les unités carbonatées ne forment pas de couches discontinues (couches sédimentaires) dans la succession volcano-sédimentaire. Au contraire, ce sont généralement des lentilles et des zones riches en carbonate (atteignant 85 m d'épaisseur) dans des roches volcano-clastiques (ce sont peut-être des brèches primaires ou des brèches tectoniques) principalement au sommet des unités rhyolitiques où elles sont recouvertes par la formation de Burwash. Le carbonate forme la matrice des conglomérats et des brèches de rhyolite associées aux dômes et coulées de rhyolite au nord du lac Sunset et des conglomérats et des brèches de gros blocs de rhyolite le long du contact volcano-sédimentaire au nord-est du lac Ross. Les principales unités se trouvent entre les lacs Turnback et Rex, près du lac 1052, au sud et à l'ouest du lac Detour. Le carbonate se présente sous forme de lentilles dans des unités de rhyolite près du lac 1052 et au contact entre la rhyolite et les andésites d'Alice entre le lac Sunset et le lac 1052. Le carbonate a été mis en place pendant les derniers stades du volcanisme ou après.

Dykes and sills. Major swarms of mafic intrusions and minor felsic dykes intrude the volcanic belts and the adjacent granitoid basement. Amphibolite 'dyke' or 'intrusion' is used here as a loose term for all diabasic to gabbroic dykes and sills metamorphosed to amphibolite grade. Amphibolite dykes and sills form: (1) dense swarms or complex multiple intrusions within the volcanic belts, (2) dense swarms within margins of the Sleepy Dragon Complex, and (3) highly deformed layers of variable attitude within the Sleepy Dragon Complex. Amphibolite layers and lenses that form intrafolial folds within granitoid gneisses of the Sleepy Dragon Complex (Davidson, 1972) are probably older than the volcanic belts.

Rare felsic dykes in the volcanic belts are associated with rhyolite-dacite dome complexes. Some felsite, aplite and granitic dykes, however, are younger than volcanism and may be related to post-deformational granitic plutons.

Undeformed and non-metamorphosed diabase and gabbro dykes that cut across volcanic, sedimentary and granitoid rocks in the area are probably Proterozoic and unrelated to the present volcanic belts.

Amphibolite intrusions within the volcanic belts parallel the general trend of the belts, internal stratigraphy, and flattening of pillow lavas. With one exception southwest of Tumpline Lake, they do not intrude the Burwash Formation. Some intruded while the volcanic belt was developing and thus form an integral part of the volcanic stratigraphy. Amphibolite dykes and sills were feeders for the pillow lavas in the various volcanic belts. Some intruded laterally before emerging as lavas.

Amphibolite intrusions form dense swarms up to 4 km wide in the Sleepy Dragon Complex east of the Cameron River belt and in granitoid gneisses along the northeastern side of the Beaulieu River volcanic belt, and extensive dykes in the Sleepy Dragon Complex along the western boundary of the volcanic belt in the Sunset Lake subarea. The general trend of these dykes is parallel to dykes in the adjacent volcanic belts and to the structurally complex boundary zones between the two terranes. In both areas shear zones separate the dyke-granitoid complexes from the volcanic belts. No amphibolite dykes have been traced across these zones.

The densest swarm of mafic dykes occurs along the northeast side of the Beaulieu River volcanic belt (Fig. 3, 35). This north-northwesterly trending swarm, about 2.5 km wide and at least 20 km long, tapers out southward but continues northward beyond the limits of Figure 3. The swarm comprises a multitude of diabasic to gabbroic dykes metamorphosed to amphibolite grade, and minor dykes of felsite, pink granite and 'fresh' diabase, separated by screens of granitic gneiss.

Dykes et filons-couches. D'importants essaims d'intrusions mafiques, et quelques dykes felsiques ont pénétré les zones volcaniques et le socle granitoïde adjacent. On emploie ici les mots « dykes » d'amphibolite ou « intrusion » de façon approximative pour désigner tous les dykes et filons-couches diabasiques à gabbroïques métamorphisés jusqu'au degré des amphibolites. Les dykes et les filons-couches d'amphibolite forment: (1) de denses essaims ou des intrusions multiples et complexes dans les zones volcaniques, (2) des essaims denses dans les marges du complexe de Sleepy Dragon, et (3) des couches très déformées de disposition variable dans le complexe de Sleepy Dragon. Les couches et les lentilles d'amphibolite qui forment des plis intrafoliaires dans les gneiss granitoïdes du complexe de Sleepy Dragon (Davidson, 1972) sont probablement plus anciens que les zones volcaniques.

Les rares dykes felsiques que l'on rencontre dans les zones volcaniques sont associés aux complexes de dômes de rhyolite-dacite. Quelques dykes de felsite, d'aplite et de granite, toutefois, sont plus récents que le volcanisme et peuvent être mis en relation avec les plutons granitiques postérieurs aux déformations.

Des dykes de gabbro et de diabase, non déformés et non métamorphisés qui coupent des roches volcaniques, sédimentaires et granitoïdes dans la région, sont probablement du Protérozoïque et sans aucun lien avec les zones volcaniques actuelles.

Les intrusions d'amphibolite dans les zones volcaniques sont parallèles à la direction générale des zones, de la stratigraphie interne et à l'aplatissement des laves en coussins. Avec une exception au sud-ouest de lac Tumpline, elles ne pénètrent pas dans la formation de Burwash. Quelques-unes ont pénétré pendant la formation de la zone volcanique et forment donc une partie intégrale de la stratigraphie volcanique. Les dykes et filons-couches d'amphibolite ont été des filons nourriciers pour les laves en coussins dans les diverses zones volcaniques. Certaines ont pénétré latéralement avant d'émerger sous forme de laves.

Des intrusions d'amphibolite forment de denses essaims atteignant 4 km de large dans le complexe de Sleepy Dragon à l'est de la zone Cameron River et dans des gneiss granitoïdes le long du côté nord-est de la zone volcanique de Beaulieu, River et de vastes dykes dans le complexe de Sleepy Dragon, le long de la limite ouest de la zone volcanique dans la sous-région de Sunset Lake. La direction générale de ces dykes est parallèle à celle des dykes dans les zones volcaniques adjacentes et aux zones limites complexes structurellement entre les deux terranes. Dans les deux régions, les zones de cisaillement séparent les complexes granitoïdes-dykes des zones volcaniques. Aucun dyke d'amphibolite n'a été suivi dans ces zones.

L'essaim le plus dense de dykes mafiques se trouve le long du côté nord-est de la zone volcanique de Beaulieu River (fig. 3, 35). Cet essaim orienté nord-nord-ouest, de 2,5 km de large environ sur au moins 20 km de long, s'amincit vers le sud, mais continue vers le nord au-delà des limites de la figure 3. L'essaim comprend une multitude de dykes diabasiques à gabbroïques métamorphisés en amphibolite et quelques dykes de felsite, granite rose et de « nouvelle » diabase, séparés par des écrans de gneiss granitique.

This dyke swarm represents a major period of multiple injection of voluminous mafic magma into granitic gneiss that had a previous history of deformation. Most dykes are multiple intrusions but, unlike sheeted dykes of ophiolite complexes, there are no extensive or continuous areas of 100 % dykes and no transitions to plutonic gabbros. The dyke swarm in the Sunset Lake subarea represents 1800 m of extension over a distance of 3000 m in a granitic gneiss terrane (i.e. 60 % extension).

It has not been demonstrated that the dyke swarm in the Sleepy Dragon Complex is the same as that in the Cameron River volcanic belt, or that the dykes are related to the volcanism that formed the belt. The significance of the mafic dyke swarms is: 1) they record extension in the basement gneisses and in the substratum upon which the volcanic belt was deposited (not necessarily at the same place); and 2) even if dykes in the Sleepy Dragon Complex are not directly related to the adjacent volcanic belt, they record an important mafic magmatic event that likely resulted in effusion of lava on the surface of sialic (continental) crust. Thus the Sleepy Dragon Complex was basement to some basaltic volcanism.

Paleovolcanic and structural interpretations. The following interpretations are idealized as a consequence of deciphering primary states after extreme deformation and metamorphism, and because of numerous structural uncertainties. Major gaps remain in the structural, geochronological and isotopic data necessary to place constraints on tectonic processes.

Volcanism in the Cameron River subarea. The earliest volcanism in this subarea was voluminous extrusion of basaltic lava as extensive subaqueous fissure eruptions (Cameron River Basalt). During each new eruption dykes intruded the preceding flows: some intruded laterally (as sills) before emerging as pillow lavas. Thus the basalt accumulated as a series of lava ridges and piles that built upwards and outwards, possibly to form an elongate shield or oceanic ridge. Within the volcanic pile, lavas ponded against fault or constructional scarps which shed talus of pillow breccia and coarse basaltic volcanic clastic material. Submarine avalanches removed portions of the pile to create broad channels or valleys which were the depository for thick lenses of volcanoclastic sediments.

Mud and silt accumulated locally during pauses in lava effusions.

Rhyolitic magma erupted through the basaltic pile from at least two centres to form lava domes flanked by aprons of talus (Fig. 52b). During eruption and destruction of the domes, subaqueous debris and mass flows distributed detritus 5-10 km from the domes into

Cet essaim de dykes représente une phase importante de l'injection multiple de magma mafique volumineux dans des gneiss granitiques déjà déformés. La plupart des dykes sont des intrusions multiples mais contrairement aux dykes stratifiés des complexes ophiolitiques, il n'y a ni vaste zone de dykes ni zone continue à 100 %, ni aucune transition vers les gabbros plutoniques. L'essaim de dykes dans la sous-région de Sunset Lake couvre 1800 m sur une longueur de 3000 m dans un terrane de gneiss granitique (c.-à-d. 60 % de la longueur).

Il n'y a pas été démontré que l'essaim de dykes dans le complexe de Sleepy Dragon est le même que dans la zone volcanique de Cameron River, ou que les dykes sont reliés au volcanisme qui a formé la zone. L'essaim de dykes: 1) permet de connaître l'extension de la zone volcanique couvrant (pas nécessairement à la même place) les gneiss du socle et le substratum; 2) enregistre un événement magmatique mafique important qui a probablement résulté de l'effusion de lave sur la surface de la croûte sialique (continentale), même si les dykes du complexe de Sleepy Dragon ne sont pas directement reliés à la zone volcanique adjacente. Ainsi, le complexe de Sleepy Dragon a servi de socle à un certain volcanisme basaltique.

Interprétations paléovolcanique et structurale. Les interprétations suivantes ne sont que des suppositions, car il est difficile de décoder des états primaires après une déformation et un métamorphisme extrêmes, et en présence de nombreuses incertitudes structurales. Il reste d'importantes lacunes dans les données structurales, géochronologiques et isotopiques nécessaires pour situer des contraintes au cours des processus tectoniques.

Volcanisme dans la sous-région de Cameron River. Le premier volcanisme dans cette sous-région s'est fait sous forme d'une extrusion volumineuse de lave basaltique par d'importantes éruptions fissurales subaqueuses (basalte de Cameron River). Durant chaque nouvelle éruption, des dykes ont pénétré les coulées précédentes: certaines ont pénétré latéralement (sous forme de filons-couches) avant d'émerger sous forme de laves en coussins. Donc, le basalte s'est accumulé sous forme d'une série de crêtes et d'amas qui grossissaient et s'élevaient, pour former probablement un bouclier allongé ou une ride océanique. À l'intérieur de l'édifice volcanique, des laves se sont accumulées contre des escarpements de faille ou structuraux qui séparent des talus de brèche en coussins et du matériel basaltique grossier volcanoclastique. Des avalanches sous-marines ont arraché des parties de l'édifice créant ainsi de vastes chenaux ou vallées où se sont déposées d'épaisse lentes de sédiments volcanoclastiques.

De la boue et du silt se sont accumulés localement pendant les pauses entre les effusions de laves.

Du magma rhyolitique a fait éruption à travers l'édifice basaltique à partir d'au moins deux centres pour former des dômes de laves flanqués de talus d'éboulis (fig. 52b). Pendant l'éruption et la destruction des dômes, des débris subaqueux et des coulées en masse ont distribué des débris jusqu'à

fault troughs and depressions in the mafic lavas and outward beyond the volcanic belt and into the sedimentary basin.

Avalanching from fault scarps in the volcanic belt produced rubble of mixed rhyolite and basalt provenance.

Major eruptions of andesitic magma (Webb Lake Andesite, Fig. 52c) followed rhyolitic volcanism. The change from basaltic to andesitic compositions was rapid. The first effusions were pillowed and massive lavas that filled elongate valleys and fault troughs in the underlying basalts. Andesitic volcanism took place in the southern part of the belt, probably while basaltic volcanism continued to the north. Following these early fissure eruptions, volcanism eventually coalesced into central eruptions.

The association of rhyolite with the beginning of andesitic volcanism suggests that following basaltic volcanism a magma source (possibly a vertically zoned chamber) was being tapped from which rhyolite differentiates were the first to erupt, followed by more mafic andesites, then finally, a return to basalt (Dome Lake Basalt) at the end of the eruptive series. The rhyolite could represent purely the end product of fractional crystallization, or magma from a new source, such as melted sial, possibly present near the site where basaltic magma was introduced to the crust.

The general lack of granitic detritus or inclusions in the volcanic succession might suggest that the volcanic pile did not develop on the granitoid basement against which it now makes contact. However, the small volume of granitic-volcanic conglomerate, quartz-feldspar arenites and quartzite of the Raquette Lake Formation, that interfingers with the volcanics near where the volcanic belt tapers out indicates proximity of a granitic source terrane at the time of volcanism.

Volcanism in the Tumpline Lake subarea. Volcanism in the Tumpline Lake subarea (Fig. 53) was strongly bimodal. In the northern areas (between Tumpline and Turnback lakes) volcanism began with the subaqueous effusions of basaltic pillow lavas. Submarine debris flows carried basaltic detritus 5-10 km from this major volcanic pile. Most of this material was deposited before the succeeding rhyolite eruptions.

In the southern parts of the subarea (south of Detour and Tumpline lakes) volcanism appears to have begun with voluminous eruption of rhyolite. Although it is possible that these rhyolites are underlain by basalts (now concealed), and the mafic pile was considerably more extensive than the present exposure (as extrapolated in Fig. 53b), the present data suggest that rhyolite was the first volcanic product. This episode of rhyolite volcanism took place

5 à 10 km des dômes, dans des fossés tectoniques et des dépressions dans les laves mafiques, et vers l'extérieur au-delà de la zone volcanique, et dans le bassin sédimentaire.

Les avalanches provenant des escarpements de failles dans la zone volcanique ont produit des produits pyroclastiques non consolidés provenant à la fois de rhyolite et de basalte.

D'importantes éruptions de magma andésitique (andésite de Webb Lake, fig. 52c) ont suivi le volcanisme rhyolitique. Le passage des compositions basaltiques à andésitiques a été rapide. Les premiers épanchements volcaniques ont été des laves en coussins et massives qui ont rempli des vallées allongées et des fossés tectoniques dans les basaltes sous-jacents. Le volcanisme andésitique a eu lieu dans la partie sud de la zone, probablement pendant que le volcanisme basaltique continuait vers le nord. Après ces premières éruptions fissurales, le volcanisme a fini par se manifester sous forme d'éruptions centrales.

L'association de la rhyolite avec le début du volcanisme andésitique suggère qu'après le volcanisme basaltique une source magmatique (probablement une chambre zonée verticalement) ait alimenté les différenciés de rhyolite qui ont été les premiers à faire éruption, suivis par des andésites plus mafiques, puis finalement, retour au basalte (basalte de Dome Lake) à la fin de la série éruptive. La rhyolite pourrait représenter simplement le produit final d'une cristallisation fractionnée, ou du magma d'une nouvelle source, par exemple du sial fondu, peut-être présent près de l'endroit où un magma basaltique a été introduit dans la croûte.

L'absence générale de débris ou inclusions granitiques dans la succession volcanique peut suggérer que l'édifice volcanique ne s'est pas développé sur le socle granitoïde contre laquelle il est en contact maintenant. Toutefois, le petit volume de conglomérats granitovolcaniques, d'arénites quartzofeldspatiques et de quartzite de la formation de Raquette Lake, qui interpénètrent les roches volcaniques près desquelles la zone volcanique s'alimente, indique la proximité d'un terrane d'origine granitique au moment du volcanisme.

Volcanisme dans la sous-région de Tumpline Lake. Le volcanisme dans la sous-région de Tumpline Lake (fig. 53) était fortement bimodal. Dans les zones les plus au nord (entre les lacs Tumpline et Turnback) le volcanisme a commencé avec des effusions subaqueuses de laves en coussins basaltiques. Des coulées de débris sous-marins ont transporté des débris basaltiques à 5 à 10 km des principaux édifices volcaniques. La plupart de ce matériau a été déposé avant les éruptions subséquentes de rhyolite.

Dans les parties sud de la sous-région (au sud des lacs Detour et Tumpline) il semble que le volcanisme ait commencé par une éruption volumineuse de rhyolites. Bien qu'il soit possible que ces rhyolites recouvrent des basaltes (masqués actuellement), et que l'édifice mafique ait été considérablement plus vaste que l'affleurement actuel (tel qu'extrapolé dans la fig. 53b), les données actuelles suggèrent que la rhyolite a été le premier produit volcanique. Cet épisode du volcanisme de la rhyolite a eu lieu à partir de

from two or three major centres stretching from the present Tumpline Lake to just south of Victory Lake. Eruption of the Sharrie Rhyolite began with explosive activity in a shallow marine environment. The explosions, possibly phreatomagmatic eruptions that broke the surface of the water, distributed pyroclastic material for several kilometres in the surrounding area, to be deposited as water-laid ash on sediments accumulating in the basin adjacent to the volcanic belt. Explosive volcanism was followed by quiet effusion of magma to form a complex of rhyolite flows and domes that possibly emerged as ephemeral islands.

In some places south of Tumpline Lake, the simultaneous eruption of basalt and rhyolite from different sources may account for intertonguing of these rock types.

Basaltic magma erupted through the rhyolite complexes and effused as massive lavas on the flanks of subaerial edifices and rhyolite-block-bearing pillow lavas that partly filled the low-lying areas of submarine topography between or on the submerged flanks of rhyolitic edifices (Fig. 53c). Basaltic eruption in the southern parts of the area may have overlapped with the major early basaltic eruptions farther north.

Along the northern and eastern sides of this belt, voluminous rhyolitic volcanism, from at least three major centres postulated to be near Rex, Turnback and Tumpline lakes, succeeded the mafic effusions. Volcanism was dominated by either shallow marine or subaerial explosive eruptions. The centre southwest of Tumpline Lake may have been a large edifice, parts of which emerged above sea level to form ephemeral islands where subaerial pyroclastic flows were preserved. Presumably, widely dispersed ash, resulting from these explosive centres, settled at various times in the adjacent seas to become interbedded with mafic and felsic epiclastic debris that accumulated on the submerged flanks of the volcanoes. Although some of the bedded rhyolites that prevail south of Tumpline Lake may represent subaqueous pyroclastic flows, presumably much of the rhyolitic detritus was distributed into the basin by sediment gravity flows (Sigurdsson et al., 1980; Schiener, 1974). Hence, part of the felsic pile probably extended beyond the limits of the underlying mafic pile to become the first volcanics deposited in that part of the basin.

Locally mafic magma erupted through the late stage rhyolites and some turbiditic sediments.

Near the end of volcanism the overall paleomorphology in this subarea probably consisted of a series of submarine ridges and mountains, some of which emerged above sea level.

deux ou trois centres principaux s'étalant de l'actuel lac Tumpline jusqu'au sud exact du lac Victory. L'éruption de la rhyolite de Sharrie a commencé par une activité explosive dans un environnement marin peu profond. Les explosions, probablement des éruptions phréatomagmatiques qui ont traversé la surface de l'eau, ont distribué des produits pyroclastiques sur plusieurs kilomètres dans la région environnante; ceux-ci se sont déposés dans l'eau sur des sédiments s'accumulant dans le bassin adjacent à la zone volcanique. Un volcanisme explosif a suivi l'effusion tranquille du magma pour former un complexe de coulées de rhyolite et de dômes qui probablement ont émergé sous forme d'îles éphémères.

À certains endroits au sud du lac Tumpline, l'éruption simultanée de basalte et de rhyolite provenant de diverses sources peut expliquer l'interdigitation de ces types de roches.

Du magma basaltique a fait éruption à travers les complexes de rhyolite et s'est épanché sous forme de laves massives sur les flancs d'édifices subaériens et sur des laves en coussins contenant des blocs de rhyolites qui remplissaient partiellement les zones basses de la topographie sous-marine entre les flancs submergés d'édifices rhyolitiques (fig. 53d) ou sur ceux-ci. Une éruption basaltique dans les parties sud de la région peut avoir recouvert les principales éruptions volcaniques antérieures plus au nord.

Le long des côtés nord et est de cette zone, un volcanisme rhyolitique volumineux, provenant d'au moins trois centres principaux supposés être près des lacs Rex, Turnback et Tumpline, ont succédé aux effusions mafiques. Le volcanisme a été dominé soit par des éruptions marines peu profondes soit par des éruptions explosives subaériennes. Le centre au sud-ouest du lac Tumpline peut avoir été un grand édifice, dont des parties émergées de la mer ont formé des îles éphémères là où les coulées pyroclastiques subaériennes ont été conservées. Probablement, les cendres très largement dispersées, résultant de ces centres d'explosion, se sont déposées à divers moments dans les mers voisines en s'interstratifiant avec des débris épicaustiques mafiques et felsiques qui étaient accumulés sur les flancs submergés des volcans. Bien que quelques-unes des rhyolites litées qui prévalent au sud du lac Tumpline peuvent représenter des coulées pyroclastiques subaqueuses, on suppose que la plupart des débris rhyolitiques a été distribués dans le bassin par des coulées gravimétriques de sédiments (Sigurdsson et coll., 1980; Schiener, 1974). En conséquence, une partie de l'édifice felsique s'étendait probablement au-delà des limites de l'édifice mafique sous-jacent et a donné les premières roches volcaniques déposées dans cette partie du bassin.

Localement, un magma mafique a fait éruption à travers les rhyolites du dernier stade et quelques sédiments de turbidité.

Près de la fin du volcanisme, la paléomorphologie globale dans cette sous-région consistait probablement en une série de crêtes et de montagnes sous-marines dont une partie émergeait de la mer.

Volcanism in the Sunset Lake subarea. Volcanism began with voluminous subaqueous effusion of basalt pillow lavas (Fig. 54). The lavas probably formed an extensive elongate hummocky field of low hills and intervening valleys reflecting an underlying fracture-fissure system. Depressions in this topography accepted detritus derived from the accumulating pile. The southern edge of this field may have ended abruptly in a steep frontal slope, whereas elsewhere the field ended gradually with a tapering apron of volcanic debris sloughing from the volcanic pile.

Magma compositions became more felsic with time, and eruptive products passed from basalt to andesite without any major break in the eruptive activity. In some areas both basaltic and andesitic lavas were erupting simultaneously, presumably from different vents.

Along with the temporal compositional change there was a change in the style and environment of eruption (Fig. 54b). Eruption changed from quiet effusion of basaltic and andesitic lavas to explosive eruption of felsic andesite, dacite and rhyolite. These volcanic products built mountains, some of which emerged above sea level, where volcanism culminated with explosive subaerial eruptions. Rhyolite erupted during the final stages of volcanism from at least three main centres.

The waning stages of volcanism saw the rise of felsic domes on these edifices. At some centres effusions of intermediate to felsic compositions overlapped and became interfingering. One dome complex (south of Sunset Lake) records a compositional transition from andesite through dacite to rhyolite. North of Sunset Lake, rhyolite volcanism began with early explosive activity followed by at least two and probably more episodes of dome growth. Near the south end of the lake, part of the felsic pile emerged above sea level where explosive eruptions deposited welded ash-flow tuffs.

Erosion degraded the stratovolcanoes of this composite volcanic pile and produced submarine fans that formed extensive aprons of volcanic detritus south of the present Payne Lake.

Deposition of the Burwash sediments mainly followed volcanism in all three subareas. No stratigraphic or structural unconformities between the Burwash Formation and the Beaulieu Group have been recognized to indicate a gap in time between volcanism and sedimentation. The sediments were derived from exposed parts of the immediate volcanic pile as well as from a granitic terrane (Henderson, 1985). That some sedimentation preceded volcanism, however, is suggested by iron-formation, metapelites and granitic cobble-bearing conglomerate at, or near the base of, the Sunset Lake Basalt, and turbidites that directly contact the southwestern side of the Sleepy Dragon Complex and locally lie between the volcanic

Volcanisme dans la sous-région de Sunset Lake. Le volcanisme a commencé par une volumineuse effusion subaqueuse de laves basaltiques en coussins (fig. 54). Les laves formaient probablement un vaste terrain allongé et irrégulier de basses collines et de vallées alternées reflétant un réseau de fissures-fractures sous-jacentes. Dans les dépressions de cette topographie, les débris dérivés des édifices d'accumulation s'entassaient. La bordure sud de ce terrain peut s'être achevée abruptement dans une pente frontale très raide tandis qu'ailleurs le terrain se mélange graduellement à un talus d'épandage conique de débris volcaniques détachés de l'édifice volcanique.

Les compositions du magma devenaient plus felsiques avec le temps, et les produits éruptifs passaient des basaltes aux andésites sans aucune rupture importante dans l'activité éruptive. Dans quelques régions, les laves à la fois basaltiques et andésitiques ont fait éruption simultanément, probablement de cheminées différentes.

Avec le changement de composition temporelle, il y a eu un changement dans le style et l'environnement de l'éruption (fig. 54b). L'éruption est passée d'une effusion tranquille de laves basaltiques et andésitiques à un type explosif d'andésite felsique, de dacite et de rhyolite. Ces produits volcaniques ont construit des montagnes, dont quelques-unes émergeaient de la mer, où le volcanisme a culminé avec des éruptions subaériennes explosives. La rhyolite a fait éruption pendant les derniers stades du volcanisme d'au moins trois centres principaux.

Les derniers stades du volcanisme ont vu la montée de dômes felsiques sur ces édifices. À certains endroits, les effusions de compositions neutre à felsique les ont recouvert et se sont interdigitées. Un complexe en forme de dôme (au sud du lac Sunset) montre une transition dans la composition allant de l'andésite à la rhyolite en passant par la dacite. Au nord du lac Sunset, le volcanisme rhyolitique a commencé par une activité explosive au début, suivie par au moins deux ou plusieurs épisodes de croissance de dômes. Près de l'extrémité sud du lac, une partie de l'édifice felsique a émergé de la mer où des éruptions explosives ont déposé des ignimbrites.

L'érosion a dégradé des stratovolcans de cet édifice volcanique composite et a produit des cônes de déjection sous-marins qui ont formé de vastes plaines alluviales de débris volcaniques au sud de l'actuel lac Payne.

Le dépôt des sédiments de Burwash a suivi principalement le volcanisme dans ces trois sous-régions. Aucune discordance stratigraphique ni structurale entre la formation de Burwash et le groupe de Beaulieu n'a été reconnue pour indiquer une lacune dans le temps entre le volcanisme et la sédimentation. Les sédiments étaient dérivés des parties qui affleurent de l'édifice volcanique immédiat ainsi que d'un terrane granitique (Henderson, 1985). Qu'une certaine sédimentation ait précédé le volcanisme, toutefois, est suggéré par une formation ferrifère, des métapelites et un conglomérat contenant des galets granitiques à la base ou près de celle-ci, du basalte de Sunset Lake et des dépôts de courant de turbidité qui sont en contact direct avec le côté sud-ouest du complexe de Sleepy Dragon et localement reposent entre la

succession and the granitoid terrane. No major facies of coarse clastics typical of continental rifts are present. Although the sparse conglomeratic and quartz-rich units may have been deposited by sediment gravity flows from distal sources, the conglomerates bearing both volcanic and granitic clasts must have been derived from source terranes, possibly fault scarps, where both rock types were exposed within the volcanic fields.

Chronicle of structural and magmatic events. Profuse mafic dyke swarms record extension in the Sleepy Dragon Complex and indicate that mafic magmatism took place through sialic crust which probably was basement to some volcanism, even if it was not that of the presently adjacent volcanic belts. The amphibolite dykes within the gneissic terrane (which had a complex deformation history prior to dyke intrusion) could well be coeval with those in the volcanic belts which are contemporaneous with volcanism.

Thus, the preferred interpretation is that volcanoes erupted through submerged sialic crust and possibly also on adjacent seafloor in one or more major areas now represented by the Cameron River and Beaulieu River belts. Sedimentation was diachronous, spanning and continuing long after volcanism.

Deformation. Horizontal shortening produced large-scale steep isoclinal folds throughout the volcano-sedimentary basin and intense strain along boundaries between volcanics and basement. In the Tumpline Lake subarea polyphase deformation produced complex fold interference patterns and large-scale isoclines and anticlinoria (Fig. 56) moulded around the Sleepy Dragon Complex. In the Sunset Lake subarea the volcanic belt deformed into a broad anticlinorium and synclinorium. The premetamorphic Amacher Granite may have controlled the position of the anticlinorium. The zircon age of 2644 ± 14 Ma (Henderson and van Breemen, pers. comm., 1986) gives a maximum age of the deformation. In the Tumpline Lake and Sunset Lake subareas some deformation of volcanics may be related to diapiric emplacement of plutons either before or during deformation.

The apparent structural and stratigraphic simplicity of the Cameron River belt is deceiving. It is proposed that the belt deformed against the Sleepy Dragon Complex and that high strain produced pseudoconformable relations within the volcanic belt and between the eastern side of the volcanic belt and the granitic basement. Thus the present volcanic assemblage represents a highly deformed belt that may be tectonically imbricated and thickened or possibly thinned.

succession volcanique et le terrane granitoïde. Aucun important faciès de roches détritiques grossières, typiques des rifts continentaux, n'y est présent. Bien que des unités conglomeratiques et riches en quartz épars puissent avoir été déposées par des coulées gravimétriques de sédiments provenant de sources distales, les conglomerats contenant à la fois les débris volcaniques et granitiques doivent provenir de terranes, probablement d'escarpements de failles, où les deux types de roches affleurent dans les terrains volcaniques.

Historique des événements structuraux et magmatiques. D'abondants essaims de dykes mafiques témoignent de l'extension dans le complexe de Sleepy Dragon et indiquent que du magmatisme mafique a eu lieu à travers la croûte sialique qui probablement a servi de socle à un certain volcanisme, même si ce n'était pas celui des zones volcaniques actuellement adjacentes. Les dykes d'amphibolite dans le terrane gneissique (qui a une histoire complexe de déformations avant l'intrusion des dykes) pourraient bien être contemporains de ceux des zones volcaniques qui sont contemporains du volcanisme.

Donc, l'interprétation préférée est que les volcans qui ont fait éruption à travers la croûte sialique submergée et probablement aussi sur le fond marin adjacent dans une ou plusieurs zones importantes, sont représentés actuellement par les zones de Cameron River et de Beaulieu River. La sédimentation était diachrone, s'étalant et continuant longtemps après le volcanisme.

Déformation. Un raccourcissement horizontal a produit des plis isoclinaux à grande échelle et à forte pente dans tout le bassin volcano-sédimentaire et une intense déformation le long des limites entre les roches volcaniques et le socle. Dans la sous-région de Tumpline Lake, une déformation multiple a donné lieu à un patron d'interférence de plis complexes et à des isoclinaux et anticlinoriums (fig. 56) qui se sont moulés autour du complexe de Sleepy Dragon. Dans la sous-région de Sunset Lake, la zone volcanique a été déformée en un vaste anticlinorium et synclinorium. Le granite d'Amacher pré-métamorphique peut avoir contrôlé la position de l'anticlinorium. La datation par le zircon (Henderson et van Breemen, comm. pers., 1986) donne un âge maximal pour la déformation de 2644 ± 14 Ma. Dans les sous-régions de Tumpline Lake et de Sunset Lake, la déformation des roches volcaniques peut être mise en relation avec l'emplacement diapirique des plutons soit avant, soit durant la déformation.

La simplicité apparente structurale et stratigraphique de la zone de Cameron River est décevante. Il est proposé que la zone a été déformée contre le complexe de Sleepy Dragon et qu'une forte déformation a produit des relations pseudoconcordantes avec la zone volcanique et entre le côté oriental de la zone volcanique et le socle granitique. Donc, l'assemblage volcanique actuel représente une zone très déformée qui probablement était tectoniquement imbriquée et épaissie ou possiblement amincie.

The three- and possibly four-phases of folding and extreme diversity of deformation patterns displayed in the Cameron-Beaulieu volcanic belts require variable stress fields. Variable boundary conditions would be imposed by the basement and pre- to syn-deformational granitic plutons. That the main deformation is Archean is indicated by ca. 2520 Ma plutons (Stockwell, 1962; Green et al., 1968; Green and Baadsgaard, 1971) of the unmetamorphosed Prosperous Lake granite that cut discordantly across the southwestern end of the Beaulieu River belt, the Sleepy Dragon Complex and the Burwash sediments. The high temperature — low pressure metamorphism accompanied plutonism and the main phase of deformation.

Late folds that show a consistent pattern in trends of axial plane foliations, regardless of the position with respect to basement, plutons and previous fold structures, record a strain of regional nature that apparently affected supracrustal rocks across the southern part of the Slave Province. This strain may have been synchronous with post-Archean wrench faulting.

The Sleepy Dragon Complex is bounded by shear zones and zones of high strain both in the granitic gneisses and the adjacent volcanic belts. Near vertical stretching lineations and kinematic indicators along boundary zones in both the Cameron River and Sunset Lake subareas, as well as fault bound slabs of basement incorporated into the Sunset Lake Basalt, are compatible with thrusting of the volcanic belts over the Sleepy Dragon Complex (Kusky, 1986a; Lambert and van Staal, 1987). In the Cameron River subarea, however, the irregular angular boundary of the Sleepy Dragon Complex does not have the appearance of a simple thrust plane. Faults in the basement that displace the contact do not seem to penetrate the volcanic belt for any appreciable distance and are not reflected in the western boundary of the volcanic belt. This irregularity could be a relic of an original faulted topography against which the volcanic belt was deformed.

The thrust scenario would infer that movement zones had shallow to moderate attitudes which subsequently must have been steepened. It has not been determined if the sense of movement is consistent all around the Sleepy Dragon Complex with thrusting, and by what mechanism attitudes were steepened to nearly vertical all around the basement terrane. Granitic gneisses in the Amacher shear zone have horizontal lineations and movement indicators suggesting dextral transcurrent movement along part of the Sleepy Dragon boundary.

The structural history in the boundary zone is obviously complex and cannot be resolved with the present data. As a first approximation it is proposed that movement and deformation in the granite-greenstone boundary zones could be related to a combination of effects such as (1) a strain gradient

Les trois ou peut-être quatre phases de plissement et l'extrême diversité des réseaux de déformations visibles dans les zones volcaniques de Cameron-Beaulieu, nécessitaient des champs de contraintes variables. Des conditions limites variables auraient été imposées par le socle et les plutons granitiques mis en place avant ou pendant les déformations. Que la principale déformation soit archéenne est indiqué par des plutons d'environ 2520 Ma (Stockwell, 1962; Green et coll., 1968; Green et Baadsgaard, 1971) de granites non métamorphisés de Prosperous Lake qui coupent en discordance l'extrémité sud-ouest de la zone de Beaulieu River, le complexe de Sleepy Dragon et les sédiments de Burwash. Le métamorphisme à température élevée et à basse pression a accompagné le plutonisme et la principale phase de déformation.

Les derniers plis qui présentent une uniformité de direction des foliations planes axiales, sans égard à la position par rapport au socle, aux plutons et aux structures plissées antérieures, témoignent d'une déformation de nature régionale qui apparemment a touché les roches supracrustales à travers la partie sud de la province des Esclaves. Cette déformation peut avoir été synchrone des failles de décrochement post-archéennes.

Le complexe de Sleepy Dragon est limité par des zones de cisaillement et des zones de forte déformation à la fois dans les gneiss granitiques et les zones volcaniques adjacentes. Des linéations d'étirement presque verticales et des indicateurs cinématiques le long des zones limites à la fois dans les sous-régions de Cameron River et de Sunset Lake, ainsi que des dalles de socle limitées par des failles incorporées dans le basalte de Sunset Lake, sont compatibles avec le charriage des zones volcaniques sur le complexe de Sleepy Dragon (Kusky, 1986; Lambert et van Staal, 1987). Dans la sous-région de Cameron River, toutefois, la limite angulaire irrégulière du complexe de Sleepy Dragon n'a pas l'apparence d'un simple plan de chevauchement. Des failles dans le socle qui déplacent le contact ne semblent pas pénétrer la zone volcanique sur aucune distance appréciable et ne se reflètent pas dans la limite occidentale de la zone volcanique. Cette irrégularité pourrait être une structure résiduelle d'une topographie faillée primaire contre laquelle la zone volcanique a été déformée.

Selon le schéma du chevauchement, les zones de mouvement auraient eu une pente légère à moyenne qui a dû s'accroître ultérieurement. Il n'a pas été déterminé si le sens du mouvement est homogène tout autour du complexe de Sleepy Dragon par rapport au chevauchement et par quel mécanisme la disposition des couches est devenue presque verticale tout autour du terrane de socle. Des gneiss granitiques dans la zone de cisaillement d'Amacher ont des linéations horizontales et des indicateurs de mouvement suggérant un mouvement de décrochement dextre le long d'une partie de la limite de Sleepy Dragon.

L'histoire structurale dans la zone limite est évidemment complexe et ne peut pas être résolue avec les données actuelles. En première approximation, il est proposé que le mouvement et la déformation dans les zones limites granite-roches vertes pourraient être reliés à une combinaison d'effets tels que (1) un gradient de déformation est prévisible là où deux

expected where two bodies of contrasting rheology are deformed together; (2) thrusting of the volcanic belt over the basement; (3) uplift of the Sleepy Dragon Complex; and (4) transcurrent movement along some boundary zones. An original irregularity along the western side of the Sleepy Dragon Complex might be preserved if the main décollement during thrusting was not at the exact contact and that movement took place along a family of imbricate shear zones within both the volcanic and granitic terranes.

Plutonism. Emplacement of plutons spanned a time probably from the late stages of volcanism to post-deformation by folding.

Some of the earliest plutons may have arched or domed parts of the volcanic pile during late stages of volcanism and possibly contributed locally to emergence of volcanic piles above sea level. Plutons of the Defeat granodiorite may, in part, predate the main folding, but continued to rise diapirically during deformation. As deformation continued, the supracrustals became progressively more tightly moulded around them. The elongate Amacher Granite in the Sunset Lake subarea appears to have controlled the position of an anticlinorium in the volcanic belt. Late post-deformational plutons (Prosperous Granite) and possibly parts of the Defeat and Meander Lake suites represent major plutonic episodes that occurred some 40+ Ma later than formation of the greenstone belt.

The highly deformed sediments and volcanics do not show a consistent regional pattern. The diverse patterns must reflect diverse stress fields and boundary conditions. Diapiric emplacement of granitic plutons may have played a role in producing the divergent strain patterns in the volcanic belts. Plutons occupy about 25 % by area of the Yellowknife basin, and near the Cameron-Beaulieu volcanic belts, pre- and syn-deformational plutons occupy major proportions of the adjacent sedimentary basin as well as the basement complexes (Fig. 38). Structural patterns in the supracrustals clearly relate to those plutons as well as to irregular basement blocks, such as the Sleepy Dragon Complex, against which the volcanics deformed. Possibly, rising granitic diapirs not only domed and steepened attitudes in sediments and volcanics, but also inflated basement blocks and contributed to their uprise. Considering the theoretical plausibility of diapiric emplacement of granitic plutons (Marsh, 1982, 1984), its implied importance in producing deformation in some volcanic arcs (Tobish et al., 1986), and the clear relationship of plutons to structural patterns in the southern Slave Province, diapirically rising plutons played an important role in modifying the deformational pattern externally imposed on the Yellowknife supracrustal basin by tectonic events.

masses de rhéologie contrastantes sont déformées en même temps; (2) un chevauchement du socle par la zone volcanique; (3) un soulèvement du complexe de Sleepy Dragon; et (4) un mouvement de décrochement le long de certaines zones limites. Une irrégularité d'origine le long du côté occidental du complexe de Sleepy Dragon pourrait être conservée si le principal décollement pendant le chevauchement n'était pas en contact exact et si le mouvement a eu lieu le long d'une famille de zones de cisaillement imbriquées à la fois dans les terranes granitiques et les roches volcaniques.

Plutonisme. La mise en place des plutons s'est étalée probablement des premiers stades du volcanisme à la phase postérieure aux déformations par plissement.

Quelques uns des plutons les plus anciens peuvent avoir arqué ou arrondi des parties de l'édifice volcanique pendant les derniers stades du volcanisme et probablement ont contribué localement à l'émersion des édifices volcaniques de la mer. Des plutons de granodiorite de Defeat peuvent en partie avoir précédé le principal plissement, mais ont continué à s'élever diapiriquement pendant la déformation. Comme la déformation continuait, les roches supracrustales sont devenues plus étroitement moulées autour d'eux. Le granite allongé d'Amacher dans la sous-région de Sunset Lake semble avoir contrôlé la position d'un anticlinorium dans la zone volcanique. Des plutons plus récents, postérieurs aux déformations (granite de Prosperous), et probablement des parties des suites de Defeat et de Meander Lake représentent des épisodes plutoniques importants qui se sont produits il y a environ 40+ Ma plus tard que la déformation de la zone de roches vertes.

Les sédiments très déformés et les roches volcaniques ne présentent pas une répartition régionale homogène. Les diverses répartitions doivent refléter divers champs de contraintes et conditions limites. La mise en place diapirique de plutons granitiques peut avoir joué un rôle dans la constitution de réseaux de déformations divergentes dans les zones volcaniques. Des plutons occupent environ 25 % de la surface du bassin de Yellowknife et près des zones volcaniques de Cameron-Beaulieu, des plutons mis en place avant et pendant les déformations occupent d'importantes proportions du bassin sédimentaire adjacent ainsi que les complexes de socle (fig. 38). Des réseaux structuraux dans les roches supracrustales sont en rapport direct avec ces plutons ainsi qu'avec des blocs irréguliers de socle tels que le complexe de Sleepy Dragon contre lequel les roches volcaniques se sont déformées. Probablement que la montée de diapirs granitiques non seulement a arrondi et rendu plus abrupte la disposition des couches de sédiments et de roches volcaniques, mais a aussi dilaté des blocs de socle et contribué à leur soulèvement. Si l'on considère la plausibilité théorique d'une mise en place diapirique de plutons granitiques (Marsh, 1982, 1984), on suppose qu'ils ont joué un rôle important dans la déformation de quelques arcs volcaniques (Tobish et coll., 1986) et que la relation entre les plutons et les configurations structurales dans la province des Esclaves est nette, les plutons ascendants sous forme de diapir ont joué un rôle important en modifiant la configuration des déformations imposées extérieurement sur le bassin supracrustal de Yellowknife par des événements tectoniques.

Tectonic models. The first models for the evolution of the Yellowknife Supergroup (McGlynn and Henderson, 1970; Lambert, 1977; Henderson, 1981, 1985) envisioned an early stage of intracratonic rifting which underwent marine inundation. Regional extension in a 3 Ga granitic basement produced horsts and grabens in which Archean supracrustal rocks accumulated for 10-15 Ma about 2670 Ma ago. Volcanoes erupted along the faulted basin margins and, subsequently, uplifted rims shed granitic and volcanic debris into the evolving basin. Rifting never extended beyond the graben stage as the basins are floored by granitic basement and there is no evidence that the pre-Yellowknife crust was ever completely rifted such that oceanic crust formed within the basins.

The model provides a logical framework of crustal fracturing followed by volcanism, subsidence and erosion to produce the sedimentary basins, and it accounts for the ubiquitous occurrence of volcanics at the angular boundary between sediments and basement complexes. However, it does not account satisfactorily for the general lack of terrestrially derived coarse clastic successions characteristic of rift sequences (Ziegler, 1982; Burke et al., 1985), the lack of expected facies changes in the sediments as they approach the granitic basement, and the absence of alkaline magmatism; nor does it provide a dynamic framework to account for the polyphase deformation and horizontal shortening ubiquitous in the supracrustal rocks. Whereas the model requires widespread granitic basement across the Slave Province, basement to the supracrustals has been documented only at a few widely spaced localities (Baragar and McGlynn, 1976).

Although a modified version of the rift model may apply to some parts of the Slave Province, it is too simple a scenario for the province as a whole. Consequently these earlier ideas have been followed by models suggesting that the greenstone belts of the Slave Province formed in back-arc basins (Kusky, 1986 a; Helmstaedt and Padgham, 1986), or represent parts of accretionary complexes overlain by trench turbidites (Hoffman, 1986; Kusky 1986 a, b, c; Fyson and Helmstaedt, 1987).

Tarney et al. (1976) first proposed that a marginal or back-arc basin provided an actualistic counterpart to Archean greenstone belts, using the analogy of the Sarmiento complex of southern Chile. Similarly, the supracrustal succession of the Yellowknife basin has been compared to this complex (Helmstaedt and Padgham, 1986) and to the Miocene Fossa Magna of Japan (Folinsbee et al., 1968). In the Cameron-Beaulieu volcanic belts, however, the ophiolitic part of this analogy is not adequately represented. A major part of the typical ophiolite sequence (Constantinou, 1980), the sheeted dykes, layered gabbroic plutons,

Modèles tectoniques. Les premiers modèles pour l'évolution du supergroupe de Yellowknife (McGlynn et Henderson, 1970; Lambert, 1977; Henderson, 1981, 1985) imaginaient qu'un premier stade de formation de fossés tectoniques intracratoniques s'est produit pendant une inondation marine. Une extension régionale dans un socle granitique de 3 Ga a produit des horsts et des grabens dans lesquels les roches supracrustales de l'Archéen se sont accumulées pendant 10 à 15 Ma, il y a environ 2670 Ma. Des volcans ont fait éruption le long des marges faillées du bassin et par la suite, ont soulevé les rebords, et déversé des débris granitiques et volcaniques dans le bassin en évolution. L'évolution des fossés tectoniques n'a jamais dépassé le stade du graben car si le socle granitique constitue le fond des bassins, il n'y a aucune preuve que la croûte pré-Yellowknife ait jamais donné naissance à un fossé tectonique, de sorte qu'une croûte océanique se serait formée dans les bassins.

Le modèle avec une fracturation crustale par du volcanisme, de la subsidence et de l'érosion amenant la formation de bassins sédimentaires semble logique et il prend en considération l'omniprésence de roches volcaniques à la limite angulaire entre les sédiments et les complexes de socle. Toutefois, il ne tient pas compte de façon satisfaisante de l'absence de successions clastiques grossières d'origine terrestre, caractéristiques des séquences de rifts (Ziegler, 1982; Burke et coll., 1985), l'absence de changements de faciès sédimentaires prévalant en approchant le socle granitique, de l'absence de magmatisme alcalin; il ne fournit pas non plus un cadre dynamique permettant de prendre en considération la déformation polyphasée et le raccourcissement horizontal présent partout dans les roches supracrustales. Alors que le modèle exige un socle granitique très étendu à travers la province des Esclaves, on ne connaît le socle des roches supracrustales qu'à certains endroits très dispersés (Baragar et McGlynn, 1976).

Bien qu'une version modifiée du modèle de rift puisse s'appliquer à certaines parties de la province des Esclaves, c'est un schéma trop simple pour la province en entier. En conséquence, ces premières idées ont été suivies par des modèles suggérant que les zones de roches vertes de la province des Esclaves se sont formées dans des bassins d'arrière-arc (Kusky, 1986; Helmstaedt et Padgham, 1986) ou représentent des parties de complexes d'accrétion recouverts par des dépôts de courants de turbidité de fossés (Hoffman, 1986; Kusky, 1986 a, b, c; Fyson et Helmstaedt, 1987).

Tarney et coll. (1976) ont proposé en premier qu'un bassin marginal où d'arrière-arc constituerait une contre-partie réelle des zones de roches vertes archéennes, par analogie avec le complexe de Sarmiento du sud du Chili. De même, la succession supracrustales du bassin de Yellowknife a été comparée à ce complexe (Helmstaedt et Padgham, 1986) et à la Fossa Magna miocène du Japon (Folinsbee et coll., 1968). Dans les zones volcaniques de Cameron-Beaulieu, toutefois, la partie ophiolitique de cette analogie n'est pas représentée de façon convenable. Une importante partie de cette séquence ophiolitique typique (Constantinou, 1980), les dykes stratifiés, les plutons gabbroïques lités et les roches ultramafiques

and tectonized layered ultramafics are missing: ultramafic rocks, although present, form sparse fault-bound slivers within boundary shear zones and could be dykes. Although there are dense mafic dyke swarms, none form a continuing layer of 100 % dykes as in true sheeted dyke complexes (Baragar et al., 1987; Gass, 1980) and the densest swarms intrude granitic gneisses. Dyke swarms as dense as those in the Cameron-Beaulieu belts prevail in high standing volcanic edifices such as the Koolau complex in Hawaii (Walker, 1986, 1987) and are probably a common feature at critical levels of pressure and stress within major volcanic edifices or areas of voluminous basaltic eruption.

The most comprehensive tectonic model yet proposed for the origin of granite-greenstone terranes is that of crustal accretion which visualizes greenstone belts as remnants of fore-arc accretionary complexes where the volcanic and subvolcanic rocks are allochthonous island arcs, aseismic ridges, submarine plateaus and seamounts (Hoffman, 1986). The juxtaposed island arcs and bathymetric highs were delaminated from subducting oceanic lithosphere and overlain by allochthonous pelagic and deep sea fan-glomerates and trench turbidites in a prograding arc-trench system. Granitoids coeval with calc-alkaline volcanism are interpreted as the accreted roots of island arcs. Older basement may represent accreted microcontinents, possibly including rift basins related to their fragmentation, or active continental margins rifted during back-arc spreading. The model accommodates a wide variety of initial environments and acknowledges that some greenstone belts were erupted on older continental crust but have been detached from their foundations.

Undoubtedly these models will be tested and modified as more structural, isotopic, and geochronological data and paleoenvironmental interpretations are obtained.

Models for the evolution of the Cameron-Beaulieu volcanic belts are constrained by the following conclusions.

1. Granitic crust was present at the time of volcanism.
2. The Sleepy Dragon Complex is about 120 Ma older than the volcanic belts.
3. The volcanoes erupted mainly in a subaqueous environment.
4. Boundaries between the Sleepy Dragon Complex and the volcanic belts are shear zones or complex zones of high strain.
5. The sedimentary basin is not presently floored by a mafic layer (McGrath et al., 1983). It could be floored by granitic crust of unknown age relative to the volcanics.

litées et tectonisées manque: lorsque des roches ultramafiques sont présentes, elles forment quelques rares écaillles limitées par les failles à l'intérieur des zones de cisaillement limites; il pourrait s'agir de dykes. Bien que les essaims de dykes mafiques soient denses, aucun ne forme une couche continue de 100 % de dykes comme dans des vrais complexes de dykes stratifiés (Baragar et coll., 1987; Gass, 1980) et les essaims les plus denses pénètrent des gneiss granitiques. Des essaims de dykes aussi denses que ceux dans la zone de Cameron-Beaulieu prévalent dans des hauts édifices volcaniques comme le complexe de Koolau à Hawaii (Walker, 1986, 1987) et sont probablement un élément commun à des niveaux critiques de pression et de contrainte dans les principaux édifices volcaniques ou zones d'éruption basaltique volumineuse.

Le modèle tectonique le plus global déjà proposé pour l'origine des terranes de granite-roches vertes est celui de l'accrétion crustale qui représente les zones de roches vertes comme des restes de complexes d'accrétion dans la partie frontale des arcs où les roches volcaniques et subvolcaniques sont des arcs insulaires allochtones, des crêtes asismiques, des plateaux sous-marins et des monts sous-marins (Hoffman, 1986). Les arcs insulaires et les hauts bathymétriques juxtaposés furent délamés de la lithosphère océanique en subduction et recouverts par des dépôts allochtones de cône alluvial, pélagiques et de mer profonde, et des dépôts de turbidité de fossé dans un système prograde de tranchées dans un scénario d'arc insulaire. Des granitoïdes contemporains du volcanisme calco-alcalin sont interprétés comme étant les racines d'arcs insulaires ajoutés par accrétion. Le socle plus ancien peut représenter des microcontinents formés par accrétion, probablement incluant des bassins de rift reliés à leur fragmentation, ou des marges continentales actives faillées en rift pendant l'agrandissement du bassin marginal. Le modèle s'adapte à une grande variété d'environnements initiaux et reconnaît que certaines zones de roches vertes ont fait éruption sur une croûte continentale plus ancienne, mais avaient été détachées de leurs fondations.

Sans aucun doute, ces modèles seront testés et modifiés au fur et à mesure que nous disposerons de plus de données structurales, isotopiques et géochronologiques et d'interprétations paléoenvironnementales.

Les modèles pour l'évolution des zones volcaniques de Cameron-Beaulieu doivent tenir compte des conclusions suivantes.

1. Une croûte granitique était présente au moment du volcanisme.
2. Le complexe de Sleepy Dragon a environ 120 Ma ans de plus que les zones volcaniques.
3. Les volcans ont fait éruption principalement dans un environnement sub-aqueux.
4. Les limites entre le complexe de Sleepy Dragon et les zones volcaniques sont des zones de cisaillement ou des zones complexes très déformées.
5. Le fond du bassin sédimentaire n'est pas actuellement une couche mafique (McGrath et coll., 1983). Il pourrait avoir pour fond une croûte granitique d'âge inconnu par rapport aux roches volcaniques.

It would appear that the Cameron-Beaulieu volcanic belts developed in an environment of submerged, extended (and possibly block faulted) continental crust. The volcanic belts and the Sleepy Dragon terrane are but a fragment of a large and complex tectonic picture which could be accommodated in either the back-arc basin or accretionary models. Assessment of the local tectonic setting, however, depends on critical observations of the granite-greenstone boundary relationships which have not been adequately determined in this study.

Magma origin and emplacement. There does not seem to be a direct analogue of the totally subalkaline volcanic assemblage of these greenstone belts (tholeiitic lavas, calc-alkaline andesite, dacite, rhyolite) to assemblages in Phanerozoic or modern cratonic rifts (characterized by alkaline magmatism). Volcanic assemblages in the Slave Province as a whole bear a similarity, in part, to those at convergent plate margins in the Phanerozoic (Condie, 1982; Condie and Hunter, 1976) and to oceanic ridge tholeiites. In the Cameron River belt, where the ratio of tholeiitic to calc-alkaline compositions is high and sialic end members are minimal, magmas are presumed to be derived from the mantle.

In the Tumpline Lake and Sunset Lakes subareas, where the proportion of rhyolite and dacite to basalt is high, melting resulting from mafic magma invading the sialic crust (Burnham, 1979; Marsh, 1984) to produce calc-alkaline magma is a more attractive model.

At the present level of erosion mafic magma moved in extensive fissure systems (individual intrusions up to 6 km long) whereas felsic magma seems to have erupted from central sources. Not all dykes necessarily reached the surface, nor did they necessarily extend vertically downwards to their source. At some level in the crust the pressure in some chambers of magma accumulation may be relieved by longitudinal flow along fractures (Vogt, 1976; Lonsdale and Spiess, 1979). If the volcanic belts are allochthonous as implied in the structural interpretation, the dyke swarms within the belts are probably not rooted beneath their present locations and are thus removed from their parent magma chambers.

Accumulation of mafic magma at the base of, or within the sialic crust, may have had enough heat to cause melting and generate the calc-alkaline and felsic magma which differentiated and erupted mainly after a profuse effusion of basaltic lavas. Felsic magma chambers presumably represent higher level accumulations of crustal melt than the mafic chambers. Abrupt changes from basalts to rhyolites and from tholeiitic to calc-alkaline suites without interruption in volcanism suggest tapping of different contemporaneous magma sources.

Il apparaît que les zones volcaniques de Cameron-Beaulieu se sont développées dans un environnement de croûte continentale submergée et vaste (et peut-être faillée en bloc). Les zones volcaniques et le terrane de Sleepy Dragon ne sont qu'un fragment d'un élément tectonique important et le complexe qui pourrait servir soit dans le modèle d'arrière-arc ou d'accrétion. L'évaluation de la mise en place tectonique locale dépend toutefois d'observations critiques des relations à la limite entre le granite et les roches vertes et qui n'ont pas été déterminées de façon convenable dans cette étude.

Origine du magma et mise en place. Il ne semble pas y avoir d'analogie directe entre l'assemblage volcanique totalement sous-alkalin de ces zones de roches vertes (laves tholéitiques, andésite calco-alkaline, dacite, rhyolite) et des assemblages dans des rifts cratoniques phanérozoïques ou modernes (caractérisés par un magmatisme alcalin). Les assemblages volcaniques dans la province des Esclaves ont une similarité, en partie, avec ceux des marges des plaques convergentes dans le Phanérozoïque (Condie, 1982; Condie et Hunter, 1976) et les tholéiites de la ride océanique. Dans la zone de Cameron River, où le rapport des composantes tholéitiques à calco-alkalines est élevé, et que les membres terminaux sialiques sont minimum, on suppose que les magmas dérivent du manteau.

Dans les sous-régions de Tumpline Lake et de Sunset Lake, où la proportion de rhyolite et de la dacite par rapport au basalte est élevée, la fonte résultant du magma mafique envahissant la croûte sialique (Burnham, 1979; Marsh, 1984) pour produire un magma calco-alkalin est un modèle plus attrayant.

Avec le niveau actuel d'érosion, du magma mafique s'est déplacé dans de vastes réseaux de fissures (intrusions individuelles atteignant 6 km de long) tandis que du magma felsique semble avoir fait éruption à partir de sources centrales. Tous les dykes n'atteignent pas nécessairement la surface ni ne s'étendent nécessairement verticalement vers leur source. À un certain niveau dans la croûte, la pression, due à l'accumulation de magma dans certaines chambres, peut être atténuée par des écoulements longitudinaux le long des fractures (Vogt, 1976; Lonsdale et Spiess, 1979). Si les zones volcaniques sont allochtones comme l'implique l'interprétation structurale, les essaims de dykes dans les zones n'ont probablement pas de racine au-dessous de leur emplacement actuel, et sont donc éloignés de leurs chambres magmatiques d'origine.

Le magma mafique accumulé à la base de la croûte sialique ou dans celle-ci, peut avoir reçu suffisamment de chaleur pour faire fondre et générer le magma calco-alkalin et felsique qui s'est différencié et a fait éruption principalement après une effusion abondante de laves basaltiques. Des chambres de magma felsique représentent probablement des accumulations plus grandes de roches corticales fondues que les chambres mafiques. Les changements abrupts, des basaltes aux volcanismes, suggèrent qu'il y a eu soutirage de différentes sources magmatiques contemporaines.

Timing of volcanism. The Cameron River and Beaulieu River volcanic belts comprise numerous deformed volcanoes that erupted through, or along, the margins of older granitoid terrane for a continuous distance of 160 km in the present map area, but extend farther to the north and east to make a total distance of 430 km. The high level of volcanic activity presumably relates to a high rate of crustal extension in these areas. The estimated original volume of the volcanic belts in the present map area is between 24 000 and 45 000 km³. This estimate is necessarily crude because of the numerous unknowns: original depth, extent and shape of the volcanic fields. The values fall well within the gravity model determined by McGrath et al. (1983) for the Yellowknife greenstone belt, suggesting that the volcanic belt is 3-5 km thick, less than 15 km wide, and roughly wedge-shaped in profile, tapering towards a sedimentary basin.

If the rate of magma supply to the volcanic belts was within the ranges calculated for Hawaii (Shaw, 1973; Swanson, 1972) or Iceland (Thorarinsson, 1967; Jakobsson, 1972), the volume estimated in the Cameron River and Beaulieu River volcanic belts could have accumulated within 50-100 000 years, assuming a constant supply rate. Lack of any recognizable time breaks in stratigraphy suggests that volcanism was more or less continuous for a time, then ceased. Even if the volume estimates are several times too low, the volcanic belts could have developed in less than 1 Ma.

INTRODUCTION

Location and access

The Cameron River and Beaulieu River volcanic belts of Archean age form a crude horseshoe pattern centred about 80 km northeast of Yellowknife, N.W.T. (Fig. 1). Most parts of the belts can be reached using light fixed-wing aircraft on floats, which are available for charter at Yellowknife.

The Cameron River system, from Fenton to Dome lakes (Fig. 2, in pocket), is navigable by boat, requiring only short portages around some sets of rapids. Numerous extensive lakes (Paterson, Webb, Upper Ross and Victory lakes) allow easy access to the central and southern parts of the belt that are not on the Cameron River drainage. The Beaulieu River system provides excellent access to the eastern and southern parts of the belt near Sunset, Turnback and Tumpline lakes (Fig. 2,3, in pocket). Although the area south of Payne Lake is studded with lakes there is no continuous major waterway that allows easy access to this part of the belt. From the major waterways, and the abundant lakes, all parts of the volcanic belt can be reached on foot. The main lakes and rivers generally are free of ice by mid-June.

Physical features

Physical features of this area are typical of the Precambrian Shield with low relief, gently rolling rocky hills and

Datation du volcanisme. Les zones volcaniques de Cameron River et de Beaulieu River comprennent de nombreux volcans déformés qui ont fait éruption à travers les marges de terrane granitoïde plus ancien ou le long de celles-ci, sur une distance continue de 160 km dans l'actuelle zone cartographiée, mais s'étendent beaucoup plus vers le nord et l'est sur une distance totale de 430 km. Le fort niveau d'activité volcanique est probablement en relation avec un taux élevé d'extension corticale, dans ces régions. Le volume original estimé des zones volcaniques dans l'actuelle zone cartographiée est d'environ 24 000 et 45 000 km³. Cet estimé est nécessairement brut à cause de nombreuses inconnues: profondeur d'origine, étendue et forme des terrains volcaniques. Les valeurs correspondent beaucoup au modèle de gravité déterminé par McGrath et coll., (1983) pour la zone de roches vertes de Yellowknife, suggérant que la zone volcanique a de 3 à 5 km d'épaisseur, moins de 15 km de large et a une forme approximative de coin (en coupe transversale), s'aminissant vers un bassin sédimentaire.

Si le débit du magma des zones volcaniques était de l'ordre de celui calculé pour Hawaii (Shaw, 1973; Swanson, 1972) ou pour l'Islande (Thorarinsson, 1967; Jakobsson, 1972), le volume estimé dans les zones volcaniques de Cameron River et de Beaulieu River pourrait être accumulé en 50 à 100 000 ans, en supposant un débit constant. Le manque d'interruption reconnaissable dans la stratigraphie suggère que le volcanisme était plus ou moins continu pendant certains temps, puis a cessé. Même si les estimations du volume sont plusieurs fois trop basses, les zones volcaniques pourraient s'être développées en moins de 1 Ma.

ridges, intervened by mazes of lakes inherited from the last glacial ice sheets. The volcanic belts tend to stand in relief above granitic and sedimentary terranes which are generally lower lying and less rugged. In some places along the Cameron River and west of Amacher Lake, the volcanic belt rises abruptly as 100 m scarps above the surrounding flat topography. These differentially-scraped glacial terrains accent geological structure so that major structural trends, volcanic belts, folds and faults, are easily traceable on air photographs.

Although the country is well wooded, glacial overburden is sparse and soil horizons are absent or poorly developed, so that most of the hills expose clean, almost continuous outcrop.

Previous investigations

J.F. Henderson and A.W. Jolliffe mapped the southern parts of the volcanic belts along the Cameron River and Beaulieu River systems at a scale of one inch to four miles during reconnaissance work in the Beaulieu map area in 1937 and 1938 (Henderson, 1938, 1939; Henderson and Jolliffe, 1941). J.B. Henderson (1976, 1985) and Henderson et al. (1972, 1973) updated the four-mile mapping in the Yellowknife and Hearne Lake (previously known as the Beaulieu River area) map areas. J.F. Henderson (1941) and Y.O. Fortier (1946, 1947) mapped the southern parts of the Cameron River Belt during detailed investigations of the Gordon Lake South and Ross Lake map areas. Geologists

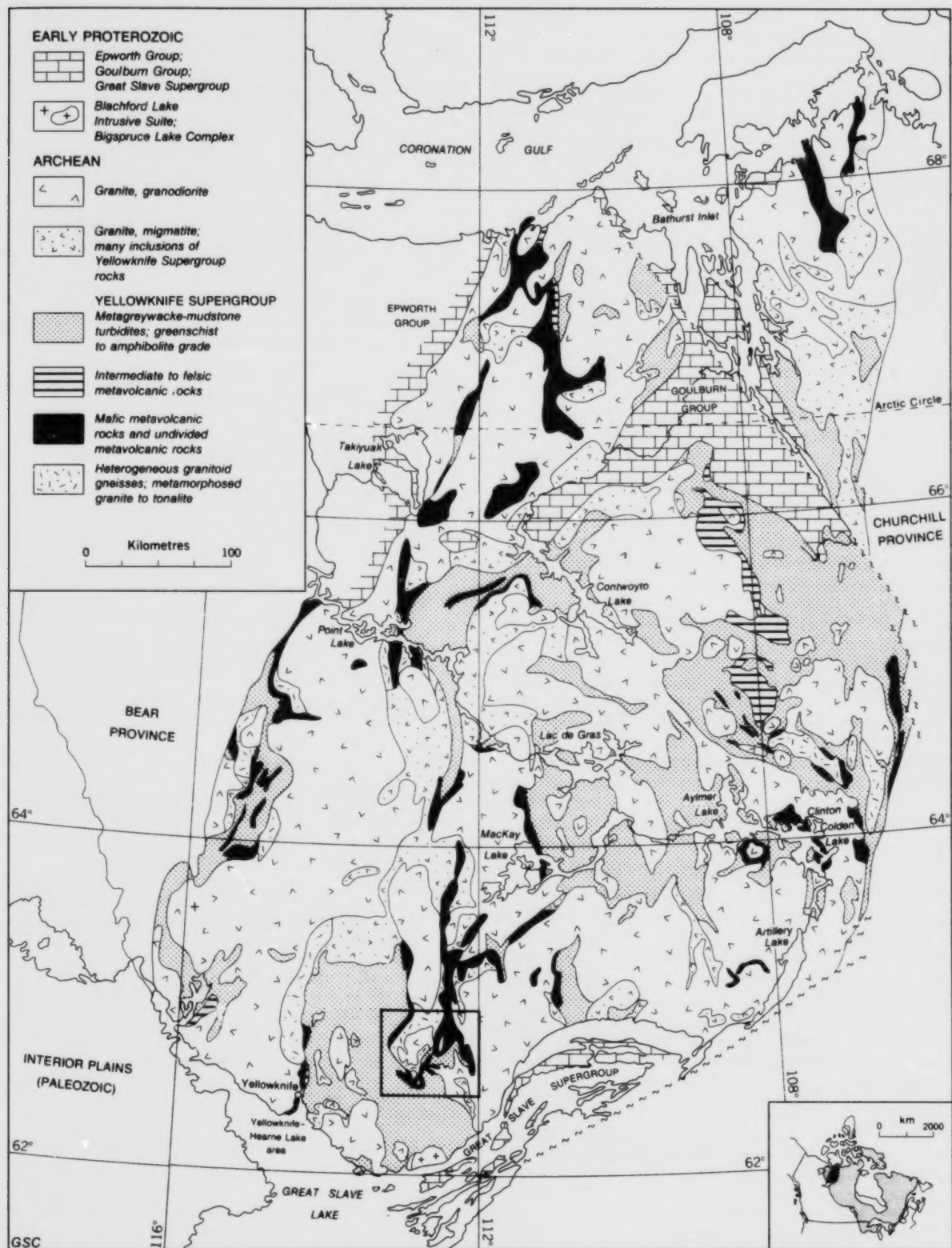


Figure 1. Generalized geology of the Slave Province (modified after Barager and McGlynn, 1976). Square shows location of the Beaulieu River and Cameron River volcanic belts.

of the Department of Indian Affairs and Northern Development have mapped parts of the northern extension of the Beaulieu River belt in the vicinity of Benaih Lake.

Although a variety of volcanic rocks (rhyolite, dacite and basaltic flows) were recognized, generally the early maps do not show any distinction within the belts. The more recent maps (Henderson, 1976; Baragar, 1966) make a two-fold division of the volcanics into dominantly mafic and dominantly felsic units.

Baragar (1966) made detailed traverses and the first systematic chemical study through a 2000 m section of the belt. He found that the acidity of the lavas increases generally with height in the lava pile and the belt (in the area sampled) contains 36 % basalt, 43 % andesite, and 21 % latite, dacite and quartz latite. Baragar (1966) first suggested that the Cameron River assemblage may rest unconformably on basement gneisses. Similar conclusions regarding the presence of granitic basement were reached by Davidson (1972) and Henderson et al. (1973).

Davidson (1972), during detailed investigations of granitic gneisses and plutons within and between the two volcanic belts, distinguished more than 17 distinct granitic units.

Fyson (1975, 1980) analyzed polyphase deformation in the Yellowknife sediments bordering the Cameron River belt and near Cleft Lake.

The southern part of the Slave Province has been prospected actively since the first discovery of gold in the late 1800s near Yellowknife. Notable are numerous occurrences of gold in shear zones within the volcanic belts and base metals near contacts between volcanic rocks and overlying sediments. Mineral occurrences near the Cameron-Beaulieu volcanic belts are plotted in Figures 2 and 3 and a summary of the economic geology of the area is presented by Henderson (1985).

Present investigations

This study began in 1972 as part of a project to update previous 1:250 000 scale mapping in the Yellowknife and Hearne Lake map areas (Henderson, 1976; 1985). It became evident early in the mapping that these belts were much more complex than the homoclinal successions of volcanic rocks previously described in the Slave Province. Consequently, the study was expanded to include a detailed examination of the volcanic belts in the northeastern parts of the Hearne Lake map area to determine: (1) internal stratigraphy, (2) the sequence and types of volcanic eruptions, and (3) the history and environment of volcanism.

The belts were mapped at a scale of 1:50 000 during the 1972 and 1973 field seasons. Fieldwork was carried out by four-man ground parties, supported by fixed-wing aircraft, and by helicopter for a two-week period in 1973.

Nomenclature of volcanic rocks follows the chemical classification of Irvine and Baragar (1971). One problem in mapping greenstone belts is the distinction of mafic and intermediate compositions amongst rocks that have been

metamorphosed variably in greenschist and amphibolite facies. Specific gravity has proved useful for this distinction amongst rocks of the Cameron and Beaulieu volcanic belts. Figure 4 shows the relationship between specific gravity and SiO_2 , normative colour index and normative plagioclase for volcanic rocks of these belts. Although none of these parameters individually correlates well with specific gravity, whole-rock compositions fall in the following general specific gravity ranges: basalt 2.8 to 3.15; andesite — 2.75 to 2.85; rhyolite and dacite — 2.68 to 2.75. Generally, where specific gravities overlap for different rock types, compositions plot near boundaries between rock types in the chemical classification.

Deformation and metamorphic recrystallization commonly change or obliterate textures critical for distinguishing fine clastic rocks as being of epiclastic or pyroclastic origin. The following terminology is used for clastic volcanic rocks whose origin is dubious. The terms are intended to be purely descriptive and nongenetic.

Volclutites are rocks containing dominantly clay-size material. They may include fine ash tuff and mudstones.

Volcarenites are rocks containing volcanic fragments of sand size and smaller, and may contain considerable amounts of clay-size material. Volcarenites may include metamorphosed equivalents of sandstone, greywacke, coarse tuffs, lapilli-tuff and fine grained hyaloclastic debris.

Rudaceous volcarenites contain cobble- and boulder-size fragments (angular or rounded) along with abundant sand-size material. These rocks include some hyaloclastites, tuff-breccias, lapilli tuffs, and pebbly wackes.

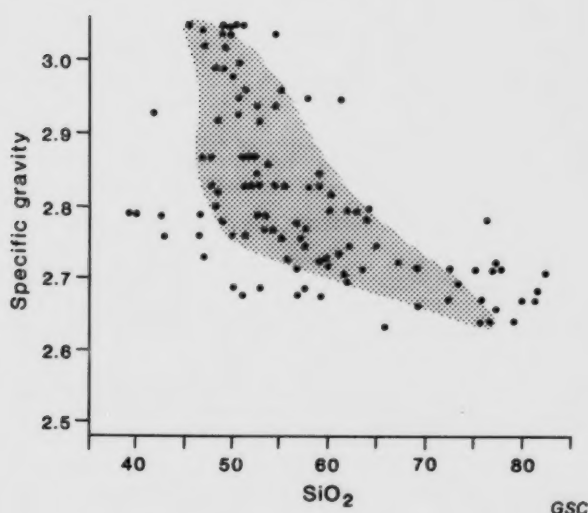
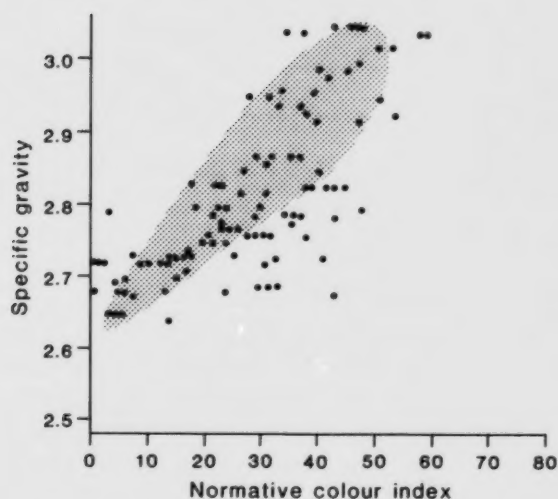
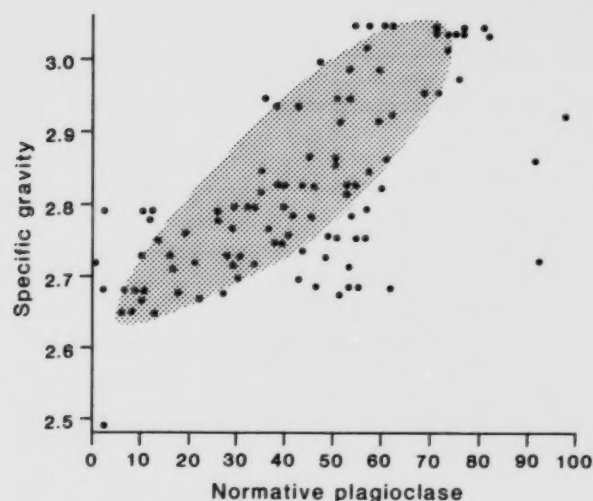
Volcrudites comprise dominantly coarse clastic material of boulder and block size. These rocks include conglomerates, avalanche breccias, pyroclastic breccias, flow-top breccias and talus debris.

The adjective **microfelsic** refers (in this report) to microcrystalline aggregates of light coloured minerals, dominantly quartz and feldspar, whose crystal size most commonly ranges between 10 and 60 microns. The aggregates contain abundant quartz in metarhyolites and metadacites, whereas feldspar may be the more abundant species (locally in absence of quartz) in meta-andesites and metabasalts.

Since all rocks discussed in this report are metamorphosed, the prefix "meta" will be dropped (from metavolcanics, metabasalts, etc.) in many cases for simplicity. The term **amphibolite intrusions** is applied to mafic dykes and sills that are metamorphosed to amphibolite grade.

Geological setting

The Cameron-Beaulieu volcanic belts lie in the southern part of the Slave Structural Province (Fig. 1), which comprises highly deformed and variably metamorphosed supracrustal rocks of Archean age that occur between extensive complexes of granitic rocks. The geology of the Slave Province, the evolution of its basins, deformation and metamorphism, have been summarized by McGlynn and



Henderson (1970, 1972), Stockwell, et al. (1970) Thompson (1978) and Henderson (1975b, 1981; 1985), and various tectonic models have been proposed by Hoffman (1986), Hoffman et al. (1986), Culshaw (1986), Grotzinger (1986), Grotzinger and Gall (1986), Kusky (1986a, b, c, in press) and Helmstaedt and Padgham (1986). Most of the supracrustal succession belongs to the Yellowknife Supergroup (Henderson, 1970) which comprises thick sequences of volcanics generally conformably overlain by greywacke-mudstone turbidites derived from a mixed felsic volcanic and granitic source. The ratio of sediments to volcanics is about 4:1. Some twenty "greenstone" belts (volcanic belts metamorphosed to greenschist and amphibolite grade), with northerly to northeasterly trends lie between sedimentary basins and vast granitic complexes. Although the volcanic belts are generally overlain by sediments, there is no evidence that volcanics extend any distance into the basin beneath the sediments (Henderson, 1981; McGrath et al., 1983). Belts in the southern parts of the Slave Province are dominantly thick successions of mafic pillow lavas with minor amounts of felsic tuffs, breccias and flows. Rare units of ultramafic rocks are known in volcanic belts near Cameron River and Point Lake (Kusky, 1986b; pers. comm.), Beaulieu River (Lambert and van Staal, 1987), and komatiitic rocks have been recognized in the Hope Bay volcanic belt in the northeastern corner of the Slave Province (Gibbins and Hogarth, 1986). Some belts in the northern and eastern parts of the province have equal amounts, or dominance, of felsic over mafic volcanics. In some areas, notably at Ross Lake (Davidson, 1972), Benjamin Lake (Heywood and Davidson, 1969) and Point Lake (Stockwell, 1933; Henderson, 1975a; Henderson and Easton, 1977), volcanics have been interpreted to lie unconformably on Archean granitoid crust.

The Cameron-Beaulieu volcanic belts (Fig. 2 and 3) are deformed around basement terrane known as the Sleepy Dragon Complex (Henderson, 1985). Both the basement and the supracrustal rocks were intruded by extensive swarms of mafic dykes and later by a series of granitic to tonalitic plutons. The Table of Formations (Table 1) applies only to the area mapped during this study.

Tectonic models proposed for the evolution of Yellowknife Supergroup vary from ensialic rifts (McGlynn and Henderson, 1970; Henderson, 1981) to back-arc basins (Folinsbee et al., 1968; Helmstaedt and Padgham, 1986; Kusky, 1986a, 1987) to suggestions that granite-greenstone terranes are allochthonous and remnants of former accretionary complexes (Hoffman, 1986; Kusky, 1986a, b, in press).

Geochronology

Geochronological data from the Slave Province are sparse (particularly in the supracrustal rocks) but suggest that the Yellowknife Supergroup volcanism took place at some time between 2650 to 2690 Ma in widely separated parts of the

Figure 4. Relationship between specific gravity and SiO_2 content, normative colour index and normative plagioclase for all rocks from the Cameron and Beaulieu volcanic belts: 95% of least altered samples plot in grey areas.

TABLE OF FORMATIONS

EON (ERA)	GROUP/COMPLEX /SUITE	FORMATION/LITHODEME	MAP SYMBOL	PRINCIPAL LITHOLOGY	
PHANEROZOIC (CENOZOIC)				Alluvium, sand, gravel	
Unconformity					
PROTEROZOIC		Indin,Milt,MacKenzie	Pdb	Diabase, gabbro dykes	
			Pf	Felsite dykes, aplite, fine grained granite	
Intrusive contact					
ARCHEAN	Plutonic Rocks	Prosperous Granite	AP	Biotite-muscovite granodiorite and monzogranite, massive, medium grained; pegmatite abundant	
		Redout Granite	AR	Biotite-muscovite granodiorite, medium-to fine-grained granite, pegmatite common: screens and inclusions of amphibolite and minor metasediment	
		Meander Lake Plutonic Suite	AML	Granodiorite, massive, medium grained biotite, minor muscovite and hornblende, concordant contacts in general; minor granite	
		Defeat Plutonic Suite	AD	Biotite granodiorite, massive, fine-to medium-grained, locally porphyritic, minor tonalite, concordant contacts with supracrustal rocks	
			AWG	Muscovite-biotite granite, massive, medium grained	
		Detour Granodiorite	ADT	Trondhjemite porphyry, leucocratic, quartz- and oligoclase-rich muscovite-biotite granodiorite and tonalite, quartz phenocrysts, fine grained	
		Amacher Granite	AAM	Biotite granite and leucocratic granite, massive, medium-to coarse-grained, minor muscovite granite	
	Intrusive contact				
			Aa	Amphibolite, dykes, sills, may include massive lavas of Ac, ATM, ASU	
	Intrusive contact				
	Yellowknife Supergroup	Duncan Lake Group	Burwash	AB	Greywacke, siltstone, mudstone, conglomerate metamorphosed to greenschist and amphibolite grade: quartz-mica (-cordierite) schists, hornfels
			Raquette Lake	ARQ	Quartzite: conglomerate and breccia containing volcanic and/or granitoid clasts
		Beaulieu Group		ACb	Carbonate, undifferentiated; within or overlying rhyolite ATs, Ar
			Cameron River Subarea *		
Dome Lake Basalt			ADO	Amygdaloidal basalt, pillow lavas, pillow breccias, hyaloclastites; coarse volcanic-clastic rocks	
Webb Lake Andesite			AW	Andesite pillowed lavas, pillow breccias: minor volcanic arenites and volc-rudites of basaltic andesite to andesite composition	
			Ar	Rhyolite; lenticular bodies, domes and associated breccias	
Cameron River Basalt			AC	Basalt pillowed lavas, pillow breccias, hyaloclastites and minor massive lava; local members of volcanoclastic rocks and lentils of volcanic arenites, breccias and rhyolite conglomerate and shale	
Tumpline Lake Subarea *					
Sharrie Rhyolite			ASH	Sparsely porphyritic rhyolite lavas, (?)domes, minor pyroclastics, crystal tuff	
Turnback Rhyolite			ATB	Felsic volcanic rocks of rhyolite to dacite composition including massive porphyritic rhyolite, tuffs, breccias and recrystallized cataclastic equivalents	
Tumpline Basalt			ATM	Pillowed lavas, minor massive lavas, layered volcanoclastic rocks	
			ATM-a	Andesite member	
Sunset Lake Subarea *					
Payne Lake			APA	Metamorphosed volcanoclastic sediments, dominantly layered schists and gneisses	
			Ar	Rhyolite, domes, flows, tuffs and volcanoclastic sediments	
Alice			AAL-d	Dacite member: dacite to felsic andesite lavas, tuff breccia; minor slate, phyllite, shale	
			AAL-a	Andesite member: felsic andesite pillow lavas, pillow breccias, hyaloclastites and minor massive lavas, pyroclastics and volcanic sediments	
Sunset Lake Basalt			ASU	Pillowed lavas, pillow breccias, hyaloclastites; minor massive lavas, iron formation and volcanoclastic sediments, polymictic conglomerate	
			AU	Ultramafic rocks, biotite schist	
Unconformity and/or intrusive contact and/or fault contact					
Sleepy Dragon Complex	(undivided)	ASD	Granodiorite, tonalite, granite ranging from massive bodies to mylonitic gneisses		

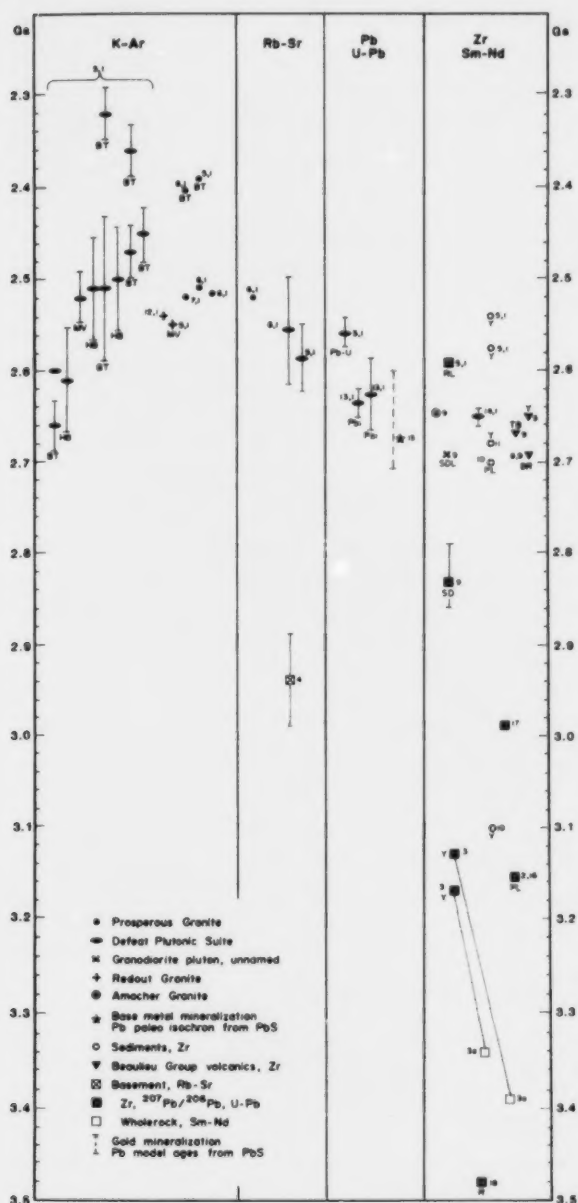


Figure 5. Radiometric dates selected mostly from units associated with the Cameron-Beaulieu volcanic belts in the southern part of the Slave Province. Vertical lines are error bars. Pbi - lead isochron, BR - Back River Complex (rhyolite dome), PL - Point Lake, RL - Ross Lake Granodiorite, SD - Sleepy Dragon Complex, TR - Turnback Rhyolite, Y - Yellowknife, R - Red Rock Lake. Numbers refer to data source: 1 - values recalculated using decay constants proposed by Steiger and Jaeger (1977) and compiled by Henderson (1985); 2 - Krogh and Gibbins (1978); 3 - Nickic et al. (1980), tonalite boulders from diatreme at Yellowknife; 3a - Bibicova et al. (in press), tonalite boulders from diatreme at Yellowknife; 4 - Frith et al. (1977); 5 - Green and Baadsgaard (1971); 6 - Green et al. (1968); 7 - Burwash and Baadsgaard (1962); 8 - Lambert and Henderson (1980); 9 - Henderson et al. (1987); 10 - Schärer and Allègre (1982); 11 - R.K. Wanless (pers. comm. 1969); 12 - Stockwell (1962); 13 - Cummings and Tsong (1975); 14 - Thorpe (1971); 15 - Thorpe (1982); 16 - Henderson et al. (1982); 17 - Frith et al. (1986); 18 - S.A. Bowering (pers. comm., 1986); 19 - Bowering and van Schmus (1984).

of $2989 \pm 6/-5$ Ma (Frith et al., 1986). The oldest $^{207}\text{Pb}/^{206}\text{Pb}$ ages of zircons from tonalitic boulders (presumably derived from basement), in a diatreme cutting volcanic rocks at Yellowknife, are 3210 and 3030-3040 Ma (Nikic et al., 1980). Whole-rock Sm-Nd dates of the same tonalite boulders are 3340 and 3390 Ma respectively (Bibicova et al., in press). The oldest known Archean basement rocks of the Bear-Slave Province are 3480 Ma (U-Pb age on zircon, S.A. Bowering, pers. comm., 1986; Bowering and van Schmus, 1984) granitoid gneisses of the Exmouth Massif in the southwestern part of the Redrock Lake area (St-Onge et al., 1984).

Basement in the Sleepy Dragon Complex on the eastern side of the Yellowknife "basin", however, appears to be only slightly older than the surrounding supracrustal rocks. Although Green and Baadsgaard (1971), reported a zircon $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2640 Ma from the Ross Lake granodiorite, zircons from granitoid gneiss at Sleepy Dragon Lake give an age of $2819 \pm 40/-31$ Ma (Henderson et al., 1987), indicating that the Sleepy Dragon Complex is about 120 Ma older than the adjacent Cameron-Beaulieu volcanic belts.

For volcanic rocks, Green and Baadsgaard (1971) reported a zircon $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2650 ± 10 Ma from dacites in the volcanic sequence at Yellowknife and a Rb-Sr isochron age for the same sequence at 2570 ± 160 Ma* ($\text{Sr}_i = 0.7022 \pm 0.0023$; recalculated from 2625 Ma). The age is similar to the Rb-Sr isochron age of 2574 ± 200 Ma* (recalculated from 2630 Ma; $\text{Sr}_i = 0.706 \pm 0.005$) that they determined for volcanic rocks in the Cameron River belt. Henderson (1981) regarded both results as minimum ages that may reflect disturbances to isotopic systems by large post-Yellowknife granitic intrusions. These data are considerably younger than the $2663 \pm 7/-5$ Ma age of zircon from the Turnback Rhyolite in the Beaulieu volcanic belt (Henderson, et al., 1987) and from a porphyry within felsic volcanic rocks of the Banting Formation in the Yellowknife volcanic belt (Padgham, 1985, p. 144). About 400 km northeast of the Cameron-Beaulieu volcanic belts, rhyolite and adjacent greywackes of the Back River volcanic complex contain zircons that give an age of 2692 ± 2 Ma

province, and sediments were deposited synchronously with volcanism or slightly later (Fig. 5).

Radiometric data from areas regarded as basement to the Yellowknife Supergroup indicate ages close to 3 Ga. At point Lake, a U-Pb concordia intercept gives an age of 3152 Ma (Krogh and Gibbins, 1978; Henderson et al., 1982). Tonalitic gneiss near the Bear-Slave boundary 300 km north-northwest of Yellowknife (Grenville Lake region) yields a Rb-Sr whole-rock isochron age of 2939 ± 51 Ma* ($\text{Sr}_i = 0.700 \pm 0.001$; recalculated from 3202 Ma: Frith et al., 1977) and zircon age (U-Pb concordia intercept age)

*Rb-Sr date recalculated using the $1.42 \times 10^{-11} \text{a}^{-1}$ rubidium decay constant.

(Van Breemen, et al., 1987). This same unit, together with detrital zircons from nearby greywackes analyzed previously, had a combined age of 2667 ± 7 Ma (Lambert and Henderson, 1980).

Zircons from greywackes at Yellowknife have a $^{207}\text{Pb}/^{206}\text{Pb}$ age of 2680 Ma (R.K. Wanless, pers. comm., 1969), and zircons from diorite boulders in a conglomerate yield ages of 2595 and 2575 Ma (Green and Baadsgaard, 1971). Single zircon grains from greywacke at Point Lake, however, gave ages either close to 2700 Ma or 3100 Ma, indicating that the sediment was derived from two sources — the granitic basement and younger volcanics or plutons (Schärer and Allegre, 1982). These data support similar conclusions by Henderson (1981).

Lead isotope data from massive sulphide deposits associated with Yellowknife, Cameron-Beaulieu, Back River, and Hackett River volcanic belts define a paleo-isochron that indicates most base metal mineralization took place at about 2670 Ma (Thorpe, 1982). Model ages for galenas from gold deposits near Yellowknife and Cameron-Beaulieu volcanic belts range from 2597 to 2705 Ma (ibid.).

Archean granites intrusive into the Yellowknife Supergroup in the Yellowknife basin in general range from 2.5 to 2.67 Ga (Green and Baadsgaard, 1971; Frith et al., 1977; J.B. Henderson and O. van Breemen, pers. comm., 1986). These plutons (see Henderson, 1985, for a comprehensive account of the plutons and their geochronology) are critical in bracketing volcanism, sedimentation, deformation and metamorphism in the Yellowknife-Hearne Lake areas into the relatively short time interval between 2667-2620 Ma (Fig. 5 and 56).

Near the Cameron River and Beaulieu River volcanic belts a small granodiorite pluton intrusive into the Sleepy Dragon Complex yields a zircon age of 2683.5 ± 2.0 Ma which is slightly older than the Turnback Rhyolite which is 2667 ± 5 Ma (Henderson et al., 1987). The pre-metamorphic (Henderson, 1985) Amacher Granite, which intrudes the Beaulieu volcanic belt, contains 2644 ± 14 Ma zircons (J.B. Henderson and O. van Breemen, pers. comm., 1986).

Widespread magmatism represented by the Defeat Plutonic Suite followed or overlapped in time with volcanism and deformation and metamorphism. Published dates for this suite by a variety of methods fall in an array between 2435-2650 Ma (Fig. 5). The younger dates, however, are by K-Ar methods and possibly reflect cooling ages or post-intrusion thermal history, whereas the older dates (2625-2650 Ma, by Pb-U and Pb isochron methods), which are close to the age of supracrustal rocks, may indicate time of emplacement (Henderson, 1985).

Accepting that the oldest dates from the Defeat Plutonic Suite represent the age of intrusion, there may have been a time gap of ca. 30-40 Ma before the next major multiple intrusive event represented by plutons of the geologically younger Prosperous Granite dated at about 2520 Ma (Stockwell, 1962; Green et al., 1968; Green and Baadsgaard, 1971). The present lack of sufficient reliable isotopic dating of plutonic episodes, however, makes this conclusion tenuous.

Acknowledgments

I extend special thanks to J.B. Henderson (GSC) for sharing field operations during the first season of fieldwork and for free interchange of ideas and continued encouragement throughout the length of this project. W.R.A. Baragar (GSC) provided unpublished field maps and original field notes from his work in the Cameron River belt, as well as chemical analyses of a transect through the belt. His thorough and conscientious critical review of the manuscript, constructive criticism, and many hours of discussion and encouragement are greatly appreciated.

Thanks are extended to T.M. Kusky for sharing results of investigations in the Cameron River subarea from the 1985 and 1986 field seasons; to C.R. van Staal (GSC) for his enthusiastic help in unravelling the structure near Amacher Lake during the 1986 field season; to R.E. Ernst (Carleton University) who shared detailed mapping of the mafic dyke swarm in the Sunset Lake subarea in 1986, and W. Davis for competent assistance and critical reading during final revisions to the manuscript.

Assistance in the field was given by P. Nadeau, R. Troyer and G. Wright in 1971; P. Geotz, A. Thomas and J. Knowles in 1973, and J. Fenton in 1986. L. Weissmann and H. Dillon-Leitch assisted in preliminary drafting, thin section photography, petrography and data compilation. W.H. Houston and J.H. Maley (GSC) provided computer programming and processing of field and chemical data. Chemical analyses were supplied by the GSC. Camp moves were by Wardair Canada Limited and Ptarmigan Airways. R. Hornel (Department of Indian Affairs and Northern Development) provided vital expediting services and radio contact in Yellowknife during field operations, as well as friendly support and generous co-operation in logistics on entering and leaving the field.

GRANITIC ROCKS

The volcanic belts wrap around a central granitic terrane whose margins correspond roughly to the Cameron River (on the west) and the Beaulieu River (on the east). Parts of this area, comprising mixed granitoid gneisses of the Sleepy Dragon Complex have been interpreted to be basement to the volcanic and sedimentary rocks of the Yellowknife Supergroup.

Plutonic rocks within and adjacent to the volcanic belts are divided into six units whose names follow the nomenclature of Henderson (1985) even though, in some cases, they do not correspond strictly to compositions determined in the present report. Data from many of the bodies are derived from their margins and may not be entirely representative of the body as a whole. Plutons are quartz-rich rocks comprising granite and granodiorite (rock types are classified following Streckeisen, 1967). Figure 6 shows the distribution and modal compositions of the various units. Modal analyses are given in Appendix 1 along with various textural and mineralogical information.

Intrusions have characteristics suggesting that they were emplaced at high or intermediate levels in the Earth's crust either during or after main episodes of regional deforma-

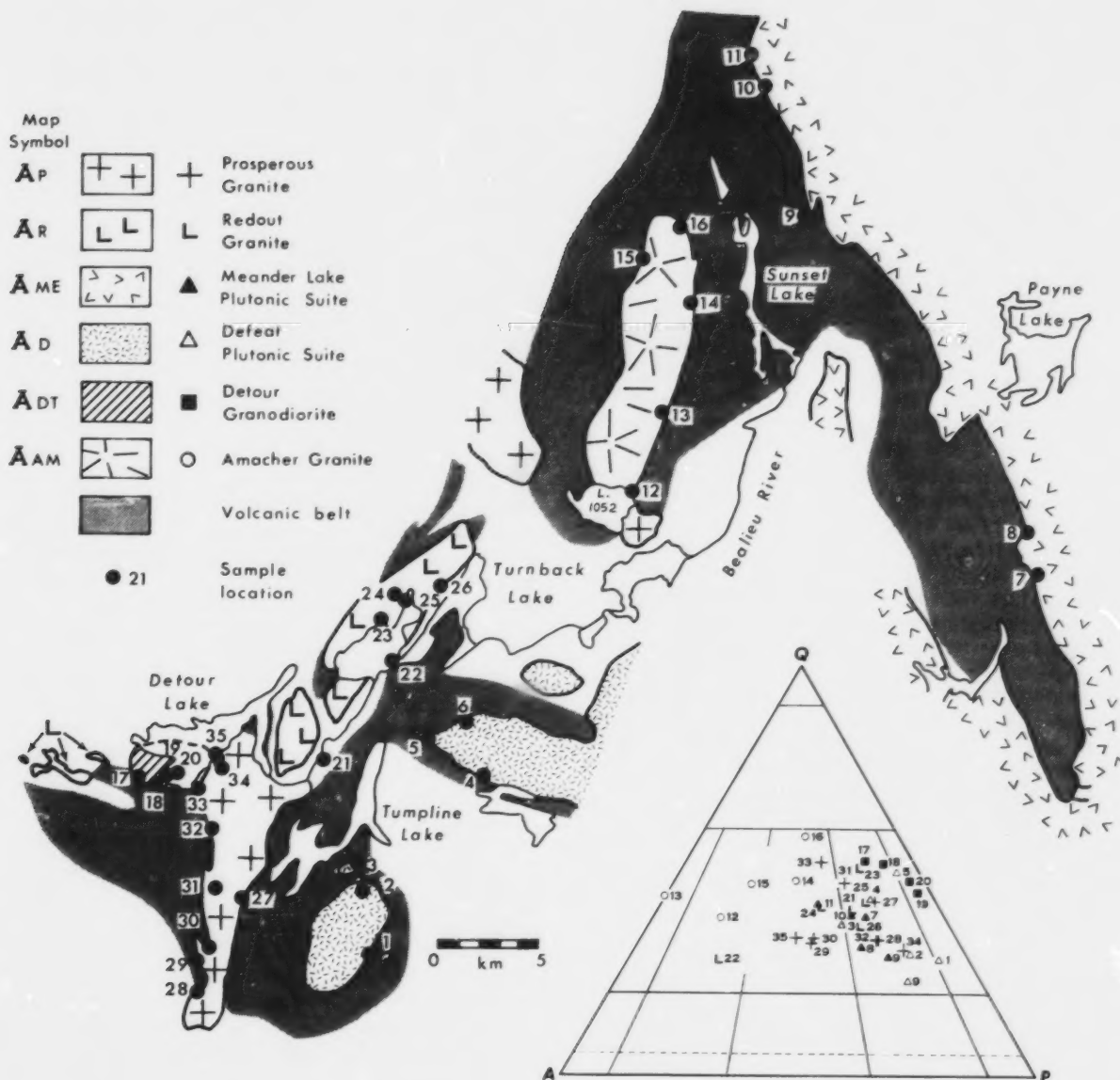


Figure 6. Modal composition of plutonic rocks near the Cameron and Beaulieu volcanic belts: Q = quartz, A = potassic feldspar and albite (An_{0-5}), P = plagioclase. Sample location numbers correspond to numbers in Appendix I.

tion. Some plutons have drastically effected the pattern of deformation in the sediments and volcanic rocks that surround them whereas others cut across all structures. Henderson (1985) considered the Detour Granodiorite and Amacher Granite to be older than the other plutons by virtue of being metamorphosed.

Sleepy Dragon Complex

The Sleepy Dragon Complex (Henderson, 1985), which forms a rectangular block between Cameron and Beaulieu rivers, comprises a metamorphosed and deformed (mostly

prior to deposition of the Yellowknife Supergroup) assemblage of mixed gneisses of diorite, tonalite and granodiorite composition that are not formally subdivided. A variety of unmetamorphosed granodiorite to granite plutons (Davidson, 1972; Henderson, 1976) that intrude the gneisses, including the Morose, Redout and Prosperous granites, are not considered part of the complex.

Davidson (1972) divided the gneissic rocks along the Cameron River belt into two varieties: (1) the Ross Lake Granodiorite (Henderson, 1941; Fortier, 1947) comprising deformed granodiorite having foliation that roughly

parallels the contact with the volcanic rocks in the vicinity of Upper Ross Lake; and (2) mixed dioritic, tonalitic and granodioritic gneisses, southeast of Sleepy Dragon Lake, having foliation with variable, folded trends that in some places is truncated by the sheared contact of the volcanic belt.

The present mapping includes only small parts of the gneissic complex where it makes contact with the volcanic belts. In several places east of the Cameron River, contacts between volcanic and gneissic terrane are clearly exposed. In most of these areas both the volcanic and granitic rocks are intensely deformed in a broad structurally complex zone near the boundary. Although there may have been an unconformable relationship between the volcanics and gneisses, the contact between the two is now sheared and faulted. At a locality 3 km north of Sleepy Dragon Lake, medium grained granodiorite gneiss with well developed lensoid streaks is mylonitized for about 10 m from the contact. The adjacent metavolcanics are highly strained for about 120 m from the contact. Near the contact, fine streaks and pods of granitic material occur within the strained amphibolite (volcanics). At the north end of Patterson Lake, lensoid, flaser structured, gneisses have mylonite layers that are themselves folded. Here foliation in amphibolite conforms to the undulatory contact but is not parallel to foliation in the gneiss. The amphibolites, in which pillows are locally identifiable, contain lenses of granitic gneiss up to 5 m wide and 8 m long. At the north end of Webb Lake, the contact between volcanic and granitic rocks is marked by a 60 m wide mixed zone dominantly of amphibolite with layers 3-5 cm thick of fine- to medium-grained granitic material. The fine grained amphibolite has subtle mineral streaking parallel to layers in the mixed zone and almost parallel to the direction of flattening of pillow lavas about 60 m from the contact. Layering in the adjacent granitoid gneiss, however, makes angles of 15-30° with layers in the mixed zone.

Where the volcanic belt pinches out to a thin unit (in some places only 20 m across) between Upper Ross and Victory lakes, previous investigators (Henderson, 1938, 1939; Fortier, 1947; Henderson, 1985; Henderson et al., 1972; and Davidson, 1972) reported conglomerate within the volcanic succession that contains both volcanic clasts and granitic pebbles identical to the Ross Lake Granodiorite. At one locality east of Upper Ross Lake Henderson (1985) interpreted a schist containing granitoid pebbles, that lies between volcanic breccia and granite, as an aluminous regolith derived from weathering of the adjacent granite with which it makes sharp contact. He likens a 0.5 m wide altered margin on the granite and some altered pebbles in the overlying metaconglomerate to weathered granite seen at the unconformity at Point Lake (Schau and Henderson, 1983).

In at least one locality near the south end of Upper Ross Lake a steeply dipping zone of tectonic breccia occurs at the edge of the granitic terrane. This granitic breccia shows all gradations eastward from coarse, non-layered and non-sorted breccia with grey carbonate matrix (15-20% of the outcrop) between blocks, through a breccia where blocks (commonly to 1 m across) are separated only by thin veins

of carbonate in fractures, to massive very fine grained granodiorite.

Along the southeastern side of the Cameron River belt a dense swarm of mafic dykes in both volcanic and granitic terranes is roughly parallel to trend of the volcanic belt. The swarm intrudes the granites for about 3 km from the volcanic belt although no amphibolite dykes have been traced across the volcanic-granitic boundary (excepting late northwest-trending dykes).

The following features have led several authors (Baragar, 1966; Davidson, 1972; Henderson et al., 1973; Baragar and McGlynn, 1976) to suggest that the Cameron River volcanic belt at one time lay unconformably on or against the gneissic granitic rocks, and that these gneisses are basement to Yellowknife Supergroup in this area: (1) the volcanic belt is a deformed homoclinal succession that in most places faces away from the granitic contact; (2) the volcanic belt truncates the trend of layering in folded gneisses near Sleepy Dragon Lake and the two units have contrasting deformational style; (3) the volcanic belt contains blocks of granitic gneiss; (4) lenses of conglomerate near the base of the volcanic succession near Upper Ross Lake contain granitic boulders identical to the Ross Lake Granodiorite which in one area could be interpreted as a regolith; and (5) the mafic dyke swarm, presumed to be intimately related to the volcanic belt, intrudes the Ross Lake Granodiorite and the volcanics, but not the Yellowknife sediments that conformably overlie the volcanic succession. Zircon geochronology (O. van Breemen, pers. comm., 1986) suggested that the Sleepy Dragon Complex is at least 30 Ma older than the adjacent volcanic belts.

Although granitic material within the volcanics could be accidentals plucked from the basement and included in dykes and lavas during eruption, or debris shed off adjacent basement, some could also represent slivers of gneiss tectonically incorporated into the highly strained contact zone.

The angular blocky outline defined by faults, having horizontal relief of 2000 m along the western side of the Sleepy Dragon Complex, that do not penetrate the volcanic belt (or where they do they have displacement much less than that of the volcanic-granitic contact), support the notion that this irregularity could be a preserved portion of an original pre-volcanic topography. Offsets of the granitic-volcanic contact along the east side of the Cameron River belt are not present at the volcanic-sediment contact along the western side of the belt. This kind of irregularity, however, is not evident along other sides of the basement block where the boundary is characterized by gently undulating sheared contacts.

Detailed investigation along a small portion of the eastern side of the Sleepy Dragon Complex west of Amacher Lake (Lambert and van Staal, 1987) showed that the contact between the granitoid basement and the greenstone cover is a complex boundary zone characterized by high strain and a mixed assemblage of basement and cover rocks. Units in the boundary zone include (1) granitoid gneisses of the Sleepy Dragon Complex, (2) ultramafic rocks, (3) Sunset Lake Basalt and (4) a swarm of diabasic and gabbroic dykes

and sills metamorphosed to amphibolites. All units trend northeasterly parallel to the general granite-greenstone boundary, and the general trend of the main foliation in the volcanic belt. The geometry and origin of this boundary zone is not well understood and geological relationships are commonly obscured by numerous amphibolite intrusions (dykes and sills).

Granitoid rocks adjacent to the volcanic belt are intensely deformed by the Amacher shear zone. Northwest of Amacher Lake two large lenticular bodies of the granitoid gneiss interleave with the volcanic-dyke complex. The southern, 800 by 150 m, lens trends parallel to the boundary zone. The northern lens thins abruptly and tapers to a 40 by 300 m tongue that cuts obliquely across the volcanic belt to the northern end of Amacher Lake. Cataclasis is present along some of the exposed contacts between these granitoids and rocks of the volcanic belt.

The southernmost lens is bound to the southeast by iron-formation and to the northwest by conglomerate. Granitoids are intensely foliated all along the southern contact which commonly is marked by a 3-5 m wide band of gossan. A thin band of iron-formation also lies between the two granitoid lenses, which are separated by a northwesterly trending dextral transcurrent fault. The granitoid gneiss lens is also cataclastically deformed where in contact with the conglomerate. This deformation and the presence of mafic dykes obscures the original contact relationships. Thus whether the conglomerate-granitoid contact is completely tectonic or a faulted unconformity has not been determined.

In at least one place west of Amacher Lake a lens of granitoid gneiss is intruded by metagabbro dykes and sills and both are deformed by the earliest deformation recognized in the volcanic belt.

A lens of conglomerate along the western side of the southern granitoid lens contains granitoid pebbles, that have a foliation that is probably predepositional. Also no intrusive relationships have been observed between the granitoids and the Beaulieu River supracrustals. These relationships suggest that the granitoid lenses either represent irregular shaped basement highs or are slices of Sleepy Dragon material that were tectonically mixed with supracrustal rocks in the boundary zone.

Along the eastern side of the Sunset Lake subarea, foliated granitoid gneisses between the Beaulieu River volcanic belt and the Meander Lake Plutonic Suite host a dense mafic dyke swarm. The gneisses form screens which make up 40% of the dyke-granitoid complex which makes fault contact (broad shear zone) with the volcanic belt. Like the Sleepy Dragon Complex, this granitoid rock has had a pre-intrusion deformational history and may represent hitherto unrecognized Archean sialic basement.

Plutonic rocks

Amacher Granite

Pink to grey weathering biotite granite and leucocratic granite form a northerly trending pluton, about 15 km long by 3 km wide, west of Sunset Lake. Abundance of quartz

(15-45%), presence of biotite, medium- to coarse-grain size, and massive nature, characterize this unit. The northern part of this body is biotite granite and has local marginal phases of muscovite granite in which the feldspar is dominantly albite. Marginal phases commonly contain about 15% quartz phenocrysts. Some rocks along the southeastern margin contain up to 25% micrographic intergrowths of quartz and microcline. These rocks, in some places, have euhedral plagioclase that is mantled by microcline.

This unit makes sharp intrusive contacts with the surrounding mafic volcanics. The pluton has a fine grained margin about 100 m wide in one area along the eastern side. Locally, along the northeastern end of the body, the granite includes large blocks of the surrounding mafic volcanics. Outcrops of this massive pluton shows no evidence of cataclasis. Thin sections from near the northern contact, however, show some granulation and recrystallization of quartz and feldspar along crystal boundaries and in fractures within the rock, indicating brittle fracture. Penetrative deformation, common in the surrounding volcanic units, however, is not evident within the pluton.

Detour Granodiorite

An elliptical stock 3 km long and 1.5 km wide is partially exposed along the south side of Detour Lake. The stock grades from tonalitic rocks near its centre to a wide (greater than 250 m) margin of muscovite-biotite granodioritic rocks. The rocks are trondhjemitic in that they are characteristically leucocratic (colour index of 4 to 8), quartz-rich (45-50%) and have plagioclase that ranges from albite to oligoclase (An_{5-15}). The stock is megascopically uniform, comprising 10-15% phenocrysts of quartz (2-4 mm across) in a fine grained, grey matrix. In thin section, quartz forms subhedral, embayed phenocrysts and the matrix (comprising anhedral oligoclase and albite commonly unwinning, microcline, biotite and muscovite) has a fine hypidiomorphic granular texture. Micas and some feldspars (averaging 0.6 mm) form microphenocrysts that tend to occur in glomeroporphyritic clusters. Accessory minerals include zircon and, locally, olive-green tourmaline.

Primary foliation in the stock is defined by the preferred orientation of biotite, tabular feldspars and glomeroporphyritic clusters of these minerals. The foliation is steeply dipping and generally parallel to the long axis of the intrusion, although it is crudely concentric and roughly parallel to the walls of the stock. There are three sets of closely spaced joints: parallel to foliation (most common), horizontal and crudely radial.

The porphyry along the western margin of the stock encloses blocks of metasedimentary material. Near the walls the blocks are not rotated and have average diameters of a few metres, whereas farther into the stock they are smaller, have been rotated, and are fewer in number. Blocks are seldom found more than 50 m from the margin.

The stock has a contact metamorphic aureole, about 20 m wide, along the southern and western side. This aureole is characterized by a slight increase in grain size (notably

of biotite) in the metasediments and by a network of quartz veinlets and lenses, generally less than 30 cm long.

The eastern rim of the stock, where it makes contact with white granite to the southeast, has a well developed granoblastic texture. This textural variation is interpreted as contact metamorphism of the trondhjemite stock by the white granite pluton.

The boundary of the stock along its western side is highly discordant to structures in the adjacent metasediments. A dyke-like apophysis of the stock, however, emerges from its northwestern corner and penetrates metasediments parallel to bedding.

Dykes of pink aplite (5-20 cm wide) and grey pegmatite (up to 100 cm wide) cut both the trondhjemite stock and the granites to the southeast.

This trondhjemite is unique amongst the other plutonic rocks in the area which are normal granites and granodiorites. The discordant nature and thin contact aureole are consistent with a high level intrusion. This pluton belongs to an early episode of plutonism, now cut by later granites, that is possibly penecontemporaneous with granites to the east of Tumpline Lake.

Defeat Plutonic Suite

Two large and three small round to oval plutons, southeast of Tumpline and Turnback lakes, are part of the Defeat Plutonic Suite which forms a widespread group of granitic intrusions in the Hearne Lake and Yellowknife areas (Henderson, 1985). In the Tumpline Lake subarea they are pink to grey, fine- to medium-grained, hypidiomorphic granular, biotite granodiorites that have 25-35 % quartz and 5 to 10 % biotite. Parts of the small (400 m across) body near the tip of the east peninsula of Tumpline Lake, however, contain 50 % quartz. Margins of the plutons, which locally are tonalite, are commonly porphyritic and contain euhedral phenocrysts of plagioclase up to 5 cm. The phenocrysts have oscillatory zoning through to their cores and myrmekite along their rims. The plutons generally are massive except for local weak preferred orientation of crystals and stringer-like clots (0.5 mm wide and 3 mm long) of biotite which define a steeply dipping foliation.

Inclusions of felsite and mafic volcanic rocks, ranging up to 10 m across, are common along the southern margin of the largest pluton southeast of Tumpline Lake. Irregular patches of the volcanics, as much as 150 m across, occur in the north-central part of this body. Similarly, angular inclusions of volcanic rocks are abundant along the northern and western sides of the large east-trending pluton. Elongate screens of volcanics within the northeastern margin of this pluton have layering that is almost parallel to that in the adjacent volcanic belt.

Layering in the surrounding volcanic and sedimentary rocks generally conforms to the outline of these plutons. In detail, however, discordant relationships are common. The contact between the large oval pluton southeast of Tumpline Lake and the adjacent volcanics is sharp. Along the southeastern side of the pluton a swarm of granite and aplitic

dykes and sills occurs in the volcanics. The smallest of the two plutons north of this body has a deformed margin. Both the margin of the pluton and the surrounding rhyolite are intensely sheared. Muscovite developed along these shear planes imparts a schistosity to both the volcanic and the granitic rocks. The other small (1100 m across) pluton in this area cuts across massive amphibolite dykes. Some aplite dykes, 10-15 cm wide, occur within the smaller pluton. Some of the granitic dykes cutting the volcanic rocks along the larger body can be traced into the pluton.

The eastern side of the east-trending pluton cuts off the volcanic belt abruptly. In this area the contact trends almost perpendicular to bedding in the volcanic succession. In detail, however, the granitic contact is very irregular and offshoots of the pluton about 1 m wide have undulatory trends where they have intruded both parallel and across bedding in the volcanic terrane. Rarely, the granodiorite of this body has a chilled margin in which the rock becomes progressively finer grained towards a sharp intrusive contact with the volcanics. Along the northwestern side of this pluton, a zone up to 150 m wide, of mixed volcanic and granitic rocks, makes a precise contact between the two units difficult to define. The mixed zone varies from tonalite choked with blocks and screens of volcanic rock to predominantly volcanic rock with a maze of dykes, sills and fine stringers to fine- to medium-grained granitic and aplitic material. In places a sharp intrusive contact is highly irregular where the granitic body has intruded the volcanics as a series of short dykes that taper out abruptly.

The following features suggest that these plutons are epizonal intrusions (Buddington, 1959): (1) their massive character with local porphyritic phases; (2) local chilled margins and narrow contact aureoles around some bodies; (3) sharp, locally crosscutting contacts between granitic and volcanic rocks; (4) abundant inclusions of volcanic rocks within the margins of the plutons; and (5) a swarm of granitic and aplitic dykes in the volcanic terrane. The presence of large felsic volcanic inclusions in the north-central part of the large pluton southeast of Tumpline Lake, suggests that the present level of erosion is close to the roof of the pluton.

The conformable nature of the gross layering around these plutons may be interpreted as due to either forceful emplacement of the plutons or to deformation of the stratified rocks around previously emplaced plutons. Either interpretation may be difficult to prove. Round outlines, strained margins, stretching lineations in enveloping volcanics that plunge steeply away from the large pluton south of Tumpline Lake, and complex structural patterns around individual bodies are compatible with diapiric emplacement (Dixon, 1975; Platt, 1980). The abundance of inclusions and crosscutting contacts, however, are evidence of stoping of the country rocks during emplacement. Regional structural patterns in the sediments and volcanics make abrupt deflections and deep embayments near these plutons and along the margins of the Sleepy Dragon Complex (see Structural Geology). A possible interpretation is that these plutons intruded the volcanic and sedimentary basin diapirically either before or during regional deformation (Folinsbee et al., 1968) and compression related to the

large-scale fold structures in the basin continued to mould the layered rocks around the plutons.

Meander Lake Plutonic Suite

This unit forms parts of large plutonic complexes that lie near Payne Lake, along the eastern margin of the volcanic belt in the Sunset Lake subarea, and south of Sunset Lake (Henderson, 1976). Near Payne Lake, the rocks comprise fine- to medium-grained, biotite-, biotite-muscovite-, and rarely, biotite-hornblende granodiorite. Pegmatite dykes are rare.

A northwesterly trending zone of intense cataclasis deforms the plutonic complex near the eastern side of the volcanic belt (Henderson, 1976). Cataclastic granite is also common along narrow northeast-trending fault zones that cut obliquely across the volcanic and granitic terrane in the Sunset Lake subarea. Cataclastic granites vary from (1) rocks that appear massive megascopically, but in thin section show laminar quartz with granulated and recrystallized margins, and broken and stress-twinned plagioclase; (2) breccias having a vague lenticular pattern of fragments; (3) rocks with megascopic fluxion structure; to (4) aphanitic ultramylonite.

The boundary between the plutonic complex and the amphibolitic volcanic complex (comprising metamorphosed mafic lavas, tuffs and dykes) is irregular. Northeast of Sunset Lake this boundary is marked by a broad (ca. 800 m wide) zone of mixed granitic and amphibolitic rocks upon which intense cataclasis is locally superimposed. In this area a swarm of elongate bodies and small (0.2 cm to 4 m wide) dykes of granodiorite intrude massive amphibolite and contain blocks of the surrounding amphibolite. Locally, however, metagabbro dykes (interpreted from massive amphibolite bodies that have coarse grained cores and fine grained margins) enclose blocks (1-15 m across) of fine grained granitic material and medium grained granodiorite. In some cases the granitic lenses are cataclastic, whereas the enclosing amphibolite is not.

Although the detailed structural relationships have not been established in this area, it appears that (1) the granodiorite plutons have intruded the volcanic terrane, and (2) a large mafic intrusive complex has intruded both the volcanic succession and the adjacent granites along the northeastern boundary between these two major map units. Furthermore, the plutonic complex, shown as many distinct bodies by Henderson (1976), probably represents several episodes of intrusion.

White muscovite-biotite granite

A body of white muscovite-biotite granite, about 1500 m long and 1000 m wide, occurs along the eastern end of the Detour Granodiorite stock. Goetz (1974) described the granite as a massive, hypidiomorphic granular, medium grained rock comprising 40% quartz, 40% potassic feldspar, 10% plagioclase, 10% biotite, and minor muscovite. That the granite is younger than the Detour stock is indicated by blocks of granodiorite porphyry enclosed in the granite and by swarms of white granite dykes that penetrate the Detour Granodiorite.

Redout Granite

Biotite-muscovite granodiorite and biotite granodiorite of the Redout Granite form two plutons west and southwest of Turnback Lake, that together have a northeasterly trend. The northern body, described in detail below, is 12 km long and 2 km wide, and the southern body is an oval 2 by 5 km. Four small irregular bodies of medium grained biotite granodiorite that intrude metasediments and metavolcanics west of Detour Lake may be related to these intrusions.

The northernmost elongate pluton comprises medium- to coarse-grained, porphyritic muscovite or biotite granite and granodiorite with granite pegmatite in the west-central and northern parts. Phenocrysts of microcline are commonly 3-5 cm long, but near the northern end of the stock they range from 10-30 cm across and lie in the medium- to coarse-grained granitic matrix. Some of the pegmatitic rocks are graphic granite. Along the northern and eastern sides the pluton has a fine grained, chilled margin, about 4 m wide, towards which the rock gradually becomes finer grained and nonporphyritic.

The pluton has incorporated abundant xenoliths of metasediment and metavolcanic material into its northern and eastern margins. Near this zone the pluton locally appears gneissic because of schlieren of partially assimilated inclusions. Of particular interest is a swarm of inclusions that lies in a band 50-300 m wide within the pluton west of Tumpline Lake. The band forms a northeast-trending inverted, U-shape that is 3000 m long and 1500 m wide. Inclusions that make up from 20 to 60% of this band vary from fine lensoid schlieren to slivers and screens ranging from 2 to 6 m wide and 10 to 30 m long. Some are 500 m long and 50 m wide. Their lithologies vary between layered felsic volcanics, layered amphibolite, biotite-hornblende schist and interbeds of the three materials. Generally, layering is parallel to the boundaries of the slivers and to the general orientation of the swarm. Although the inclusions are discrete bodies within the granite, they all have the same orientation (invariably steeply dipping) in any one area, and their lithologies, where inclusions are viewed as a group, outline the relict stratigraphy of the included formations; this is true along the western side of the U-shaped swarm.

The boundary of the pluton generally follows the trend of stratigraphy in the country rocks and locally along the eastern margin the relationship is concordant. In many places, however, this contact is very irregular and in detail it cuts across layering in the metasediments and metavolcanics.

In one place near the northwestern boundary of the pluton blocks of granite (50-250 cm across) occur in a mylonitic felsic matrix. Although the precise relationship here was not determined, a fault zone may mark the contact of the pluton.

A swarm of small (40-500 cm across) irregular to elongate granitic masses intrudes the volcanic and metasedimentary rocks to form a halo less than 500 m wide around the northern and northeastern sides of the pluton. Elongation of these bodies tends to follow the stratigraphic trends of the enclosing layered rocks. In these areas numerous granitic dykes and sills intrude the surrounding rocks and locally

pegmatite dykes (less than 1-2 m wide) criss-cross the contact zone, cutting both sediments and granite. Some quartz veins, associated with pegmatite and granitic dykes, contain disseminated molybdenite.

A contact metamorphic aureole around the pluton is recognized by (1) a slight coarsening of grain size and enrichment of muscovite in felsic volcanics that produces a sugary texture and schistose structure; (2) mica enrichment in greywackes and shales, and the appearance of garnet-cordierite-biotite assemblages in pelitic rocks along the eastern side of the pluton; and (3) very coarse grained, garnet-diopside skarn zones (7 m wide) adjacent to small intrusions into carbonate rocks off the northeastern end of the pluton. Silicified metasediments and metavolcanics near the north-western corner of Turnback Lake contain pyrite, chalcopyrite and magnetite.

The intrusive nature of this pluton is evinced by the contact metamorphic aureole, abundant inclusions of country rock in the margin of the pluton, apophyses and dykes of granite cutting the country rocks, and sharp, crosscutting contacts between granitic and layered country rocks. These features are consistent with epizonal plutons. The unit as a whole probably represents multiple intrusion and the U-shaped swarm of inclusions may represent relics of a fold that was caught between two intrusive phases of the pluton. Elongation of the plutons suggests that their emplacement was influenced by the northeast-trending regional structural pattern. This relationship and the presence of a fold-shaped relic of mafic volcanics that is riddled with small granitic bodies, along the southwestern side of the intrusion, suggest that the emplacement of the pluton occurred after a main period of regional deformation.

Prosperous Granite

Biotite-muscovite granodiorite and granite of the Prosperous Granite form three highly discordant plutons in the area mapped: (1) a north-trending body 15 km long and 1 to 5 km wide, west of Tumpline Lake; (2) an irregular body about 2 km across on the east side of Lake 1052; and (3) a larger body 2.5 km northwest of Lake 1052. This granite has been correlated with the Prosperous Granite at Prosperous Lake by Henderson (1985).

The pluton west of Tumpline Lake is a composite intrusion (the following description refers to this body only). Generally, the rocks are pink to grey, medium grained biotite- and muscovite-bearing granodiorite. Along the southwestern margin of the pluton, however, the rock is a porphyritic monzogranite that contains about 20% phenocrysts of tabular microcline up to 4 cm long in a medium grained, hypidiomorphic granular matrix. Accessory minerals include apatite, magnetite and rarely, garnet. At the northern end the pluton varies from a white weathering, massive, muscovite-biotite monzogranite to pink weathering, foliated biotite-muscovite granodiorite.

Locally, preferred orientation of micas, and in some places of tabular feldspars, define a steeply dipping foliation that is almost parallel to the contact with surrounding rocks. Although there is no evidence of cataclasis in outcrops,

quartz having pronounced undulatory extinction, fractured and bent plagioclase having curved, tapered, twin lamellae, and bent cleavage in muscovite, indicate strain.

Inclusions of country rock are abundant all along the margins of the pluton. Near the southern end, the pluton encloses huge areas of felsic volcanic rock up to 1700 m long and 400 m wide, as well as abundant smaller blocks, elongate slivers and screens. One such sliver is 18 m wide by 500 m long. Some elongate inclusions lie parallel to a steeply dipping mica foliation in the granite. Southeast of Detour Lake white weathering monzogranite includes tabular blocks of trondhjemite porphyry up to 30 m long. These inclusions, which occur up to 500 m from the contact, are recrystallized and contain poikiloblastic microcline and corroded muscovite.

Contacts between granite and the surrounding rocks are invariably sharp and in most places cut across layering in the volcanics. Locally, near the southwestern end, the rock becomes finer grained towards the contact, but no chilled margins were observed. Abundant muscovite and microcline present in the felsic volcanics at some contacts with granite define a contact aureole.

Along the southwestern margin of the pluton, sills of granite up to 7 m wide invade the felsic volcanics. Near contacts of sills, quartz veins are common along the schistose foliation in adjacent volcanic rocks. Dykes of white granite cut through the trondhjemite stock at Detour Lake.

This pluton has intruded the surrounding volcanic and sedimentary rocks and some granitic rocks at the northwest end. At the southern end of the pluton, sharp contacts, abundance of large inclusions, and slivers and masses of country rock across the width of the pluton, suggest that the present level of erosion may be near the roof of the intrusion. The highly discordant nature of this body, compared with the other plutons, and its intrusive nature into other "granites", suggest that this may be one of the youngest plutons in the area mapped.

BEAULIEU GROUP

The Beaulieu Group contains most of the volcanic belts in the southern parts of the Slave Province (Henderson, 1970, 1985).

In the areas mapped the volcanic belts are divided into three subareas, namely, (1) Cameron River subarea, (2) Tumpline Lake subarea and (3) Sunset Lake subarea (see Fig. 2, 3, in pocket). The Cameron River subarea contains the volcanic belt between Fenton Lake and Upper Ross Lake. The Tumpline Lake subarea includes the southern parts of the volcanic belt, centred on Tumpline Lake, that wraps around the Sleepy Dragon Complex from Sharrie Lake in the west to the north end of Turnback Lake. The Sunset Lake subarea contains the eastern parts of the volcanic belts along the drainage of the Beaulieu River.

Each subarea has a unique stratigraphy and style of deformation. Legends of Figures 2 and 3 show schematic facies relations (notwithstanding deformation in the three subareas) and Figure 7 shows the relative proportions of the different volcanic lithologies.

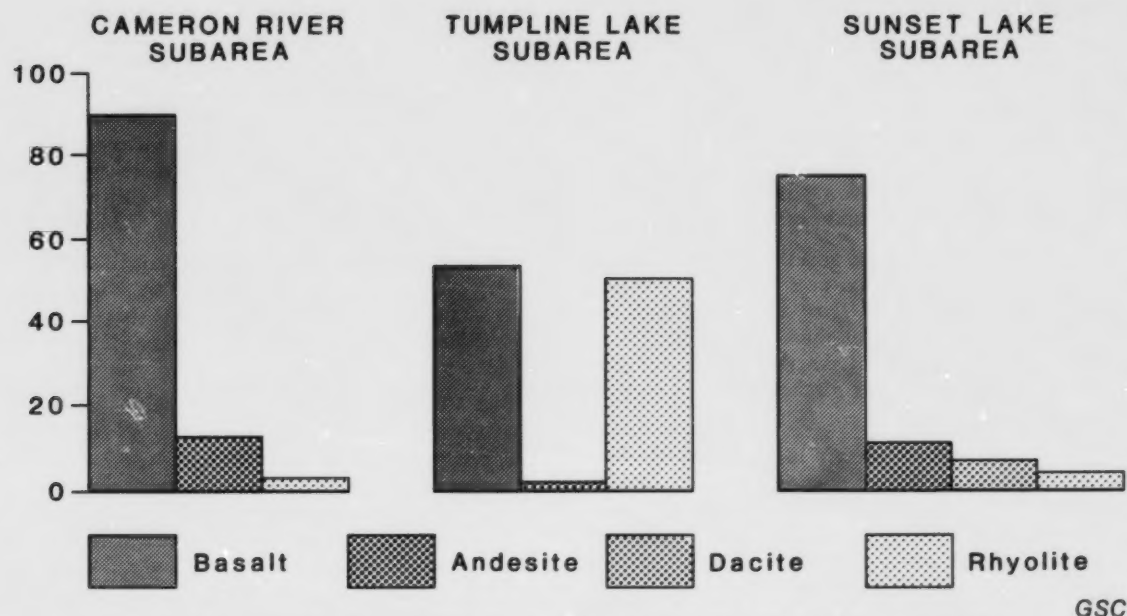


Figure 7. Proportions of volcanic lithologies in the Cameron River, Tumpline Lake and Sunset Lake subareas. Proportions represent aerial distribution of lithologies computed from 1:50 000 maps. No attempt is made to estimate volumes.

Although each subarea has a different stratigraphy, generally volcanism began with voluminous effusion of mafic pillow lavas and ended with various amounts of rhyolite. The volcanic succession probably developed penconemporaneously in all three subareas.

Cameron River subarea

The Cameron River subarea (Fig. 2, 8, in pocket) contains 40 km of a northerly trending volcanic belt (the belt continues for another 45 km north of the area mapped) that marks the boundary between sedimentary rocks of the Yellowknife Supergroup to the west and granitic rocks of the Sleepy Dragon Complex to the east (Fig. 1). The belt trends southwesterly between Fenton and Webb lakes but turns near Dome Lake to a southeasterly trend. Width of the volcanic belt varies from about 4200 m east of Fenton Lake to 3000-3500 m between Allan and Webb lakes, then tapers abruptly to about 50 m north of Upper Ross Lake, and pinches out near Victory Lake. Stratigraphic thickness of the belt is difficult to establish because (1) the belt may represent a series of flows that overlap progressively westward, but never were a thick vertical pile, (2) the belt is highly deformed and present data cannot resolve the degree to which tectonic flattening, stretching and possibly fault imbrication have altered the original thickness, and (3) the pile was expanded by intrusion of a huge swarm of dykes and sills. Some of these sills, however, intruded while the volcanic belt was developing and thus form an integral part of the volcanic succession. No attempt is made to compensate for the dyke swarm in calculations of thickness. Thus Figure 8 presents a somewhat idealized stratigraphy of the belt. The complexity of units in the central parts of the belt

(Fig. 2), compared to the northern and southern parts, reflects variability in the detail of mapping.

The eastern side of the belt is irregular where it makes faulted or sheared contacts against granitic rocks of the Sleepy Dragon Complex. The present horizontal relief of this contact is about 2000 m. In contrast, the western side of the belt is smooth and gently undulating except southwest of Webb Lake where it makes three, 300-500 m protuberances.

The volcanic belt generally is a westward-facing homoclinal succession of rocks comprising four formations:

1. Cameron River Basalt — dominantly pillow lavas of basaltic composition, 85 % of the belt;
2. Rhyolite to dacite domes, flows, breccias and arenites, 3 % of the belt;
3. Webb Lake Andesite — pillow lavas, breccias and volcanoclastic rocks, 10 % of the belt; and
4. Dome Lake Basalt — lavas and volcanoclastic rocks, 2 % of the belt.

Although the units are believed to be younger westward (toward the top of Fig. 8a) this succession does not necessarily represent a vertically stacked pile. Figure 8b shows detailed sections across the volcanic belt.

The Raquette Lake Formation is a small but significant epiclastic unit that lies between the volcanic belt or the Sleepy Dragon Complex and the Burwash Formation sediments between Upper Ross Lake and Victory Lake.

Cameron River Basalt

Definition, thickness and contact relations

The Cameron River Basalt consists dominantly of basaltic pillow lavas, pillow breccias and minor massive lavas. East of Allan and Fenton lakes the formation contains a volcanoclastic member comprising basaltic breccias, pillow breccias and basaltic arenites, and lentils of rhyolite boulder conglomerate, breccia and arenites, and lentils of shale.

Near the northern part of the area the formation makes up the entire belt. Apparent thickness is 2000 to 3300 m between Fenton and Patterson lakes and 1000 to 2000 m near Webb Lake. The formation thins abruptly to 700 m about 4 km north of Upper Ross Lake, then tapers gradually to 50 m near the lake, and finally pinches out near the south end of the lake.

The Cameron River Basalt is structurally unconformable against the Sleepy Dragon Complex. Generally volcanic rocks along this boundary are intensely sheared or the contact is a fault. The contact between the Cameron River Basalt and the Burwash Formation east of Fenton Lake is conformable. At one place between columns 10 and 11 (Fig. 8b) sediments of the Burwash interfinger with the volcanics over a distance of about 100 m. Between Fenton and Allan lakes the Cameron River Basalt is conformably overlain by the Webb Lake Andesite. West of Webb Lake, Cameron River Basalt interfingers with Webb Lake Andesite over an apparent stratigraphic thickness of 1400 m. The southern side of this zone of interfingering is in part fault-bounded against the Cameron River Basalt. Precise contact relations in this zone of interfingering are obscured by a swarm of thick mafic dykes. Large rhyolite bodies overlie the Cameron River Basalt south of Webb Lake.

Stratigraphy

Figure 8 shows stratigraphic relations of the various members and lentils within the Cameron River Basalt. Amphibolite dykes and sills, locally making up as much as 35% of parts of the formation, are discussed separately at the end of this section.

Mafic lavas and related rocks. The Cameron River Basalt contains numerous simple to compound, flow units. Individual units may be distinguished by:

1. Differences in lithology from adjacent lavas — aphanitic pillowed amphibolite, very dark green basalts, and feldsparphyric amphibolite;
2. Repeating upward cycles comprising closely packed pillow lavas to pillow breccias to hyaloclastites; and
3. Interflow units of fine grained volcanoclastic arenites (meta-hyaloclastites that divide thick monotonous successions of pillow lava) shale, siltstone, rhyolite fragment conglomerates and breccias.

Units of pillow lava and pillow breccia generally range from 20 to 80 m thick. Uninterrupted successions of pillows, which cannot be subdivided on the basis of structure or lithology, in some places range up to 250 m thick, but

are generally less than 100 m. In many places closely packed pillows grade upward into pillow breccia and hyaloclastic rudites and arenites or mixtures of these rocks. This succession is commonly overlain by more close-packed pillow lavas. At one locality, however, this upward changing succession is the exact reverse: a unit, 25 m thick, grades upwards from basal volcanoclastic breccias through pillow breccias to pillow lava. The pillow lava that forms the top of the unit is overlain conformably by siltstones.

Locally the upper 100 m of the Cameron River Basalt, where it makes contact with overlying Webb Lake Andesite, is broken pillow breccia containing closely packed, irregular, angular, fragments (5-7 cm across) of amygdaloidal basalt and the occasional pillow.

Pillows within closely packed successions generally conform tightly to one another and have no hyaloclastite between them (Fig. 9). On horizontal surface they appear as undeformed equant sacks to irregular shapes, ranging from 15 to 180 cm (and locally to 250 cm). Long tabular shapes are rare on horizontal surfaces. In many areas, however, pillows are elongate vertically and ratios of length to breadth of 4:1 are common (Fig. 10). That the elongation

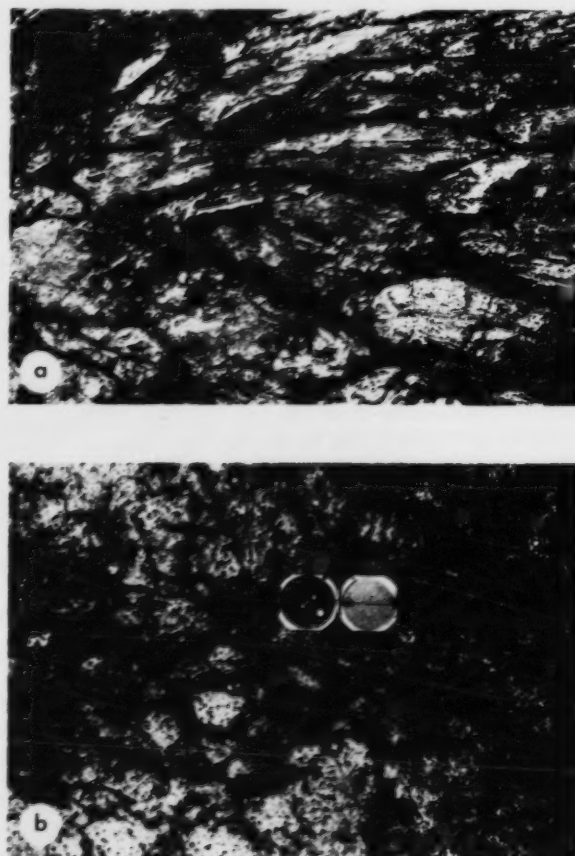


Figure 9. (a) Close-packed pillow lavas in the Cameron River Basalt. GSC 170716 (b) Pillow breccia in Cameron River Basalt. GSC 170820



Figure 10. Tectonically deformed pillows and breccia in the Cameron River Basalt east of Allan Lake: (a) Steeply plunging elongate pillow lavas. GSC 170708 (b) Tectonically elongated fragments in deformed breccia plunge 85°. GSC 170672

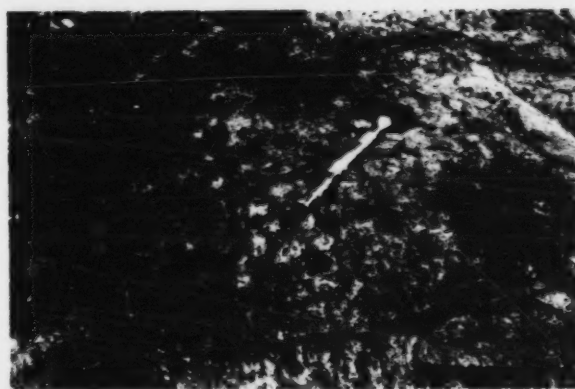


Figure 11. Thin selvages and vesicle zones in pillow lava of the Cameron River Basalt west of Webb Lake. Selvages 5 to 8 mm thick stand in relief. Zones of abundant vesicles are 40 to 150 mm wide. Vesicles range from 1 to 15 mm but generally are less than 5 mm. GSC 170715

of pillows may have been influenced by tectonic stretching or flattening is suggested by the constant orientation of elongation over broad areas and tapering of pillows near their ends when seen in vertical section.

Relics of glassy selvages on pillows are very dark green weathering, aphanitic micro-amphibolite. They commonly range from 2 to 20 mm thick and are up to 30 mm in pillows with unusually thick rims. Selvages commonly weather in recession and clearly mark the outlines of individual pillows. Locally the precise contact between two pillows is marked by a thin ridge standing in relief above the recessive selvages on either side (Fig. 11). In many places unusually large pillows tend to have irregular shapes and thick rims.

Feldsparphyric lavas (and related dykes and sills), a distinct lithological variation from the aphyric amphibolite that

makes up most lavas, occur midway through the formation east of Allan Lake and within the upper third of the formation in two localities southeast of Webb Lake (columns 3 and 5, Fig. 8b). The approximate thickness of these lavas is 300 m east of Allan Lake, 275 m and 160 m near sections 5 and 3 (Fig. 8) respectively. At these localities, pillow lavas make up 25 to 40% of the porphyritic rocks. Although the lateral extent of these units is not known, the unit east of Allan Lake has been traced for about 1000 m. In column 5 (Fig. 8), a 140 m succession of fine grained pillow lavas grades upwards into 20 m of slightly coarser grained massive lavas; both units containing sparse feldspar phenocrysts as well as blebs of quartz. According to W.R.A. Baragar (unpublished information) there is no recognizable boundary between the two units except for the appearance of pillows.

Massive dolerite to metagabbro, 40 to 60 m thick, (columns 3 and 5, Fig. 8) containing sparse feldspar phenocrysts from 2 to 5 cm across, occurs at the base or within feldsparphyric pillow lavas. Pillows are almost as coarse grained as the gabbroic bodies. These gabbros are interpreted as sills or dykes (contacts between gabbros and pillow lavas are commonly sheared) but they could be lavas. One gabbroic body can be traced for about 1500 m. In several localities feldsparphyric lavas are overlain by breccias, conglomerates or grits containing rhyolite fragments.

Massive lavas make up less than 5% of the formation. Generally they are fine grained to aphanitic amphibolite that is structureless and lacks contact features common in sills and dykes such as fining of grain size towards the margins. In one place (section 5, Fig. 8) vesicular zones mark the contact between massive lava and overlying pillow lavas. In another place massive lava grades into basal flow breccia. Where contact relations are not clearly exposed, it is commonly impossible to distinguish massive lavas from thin, fine grained amphibolite sills or dykes. Units of massive lava that have been distinguished range from 6 to 20 m thick.

Volcanic breccias, interpreted as hyaloclastites, that separate pillowed and massive lavas, commonly vary from 10 to 50 m thick and rarely range up to 250 m thick. They make up between 1 and 38% of most sections (column 10, Fig. 8b) and average about 5 to 10%. These units are massive to vaguely bedded, although some have well developed thin bedding. In contrast, local epiclastic mafic units are well bedded and locally contain fragments of rhyolite. Most units are local pods and lenses that are not extensive laterally. Correlation of units is simplified by removing large sills and dykes from the stratigraphic succession (and thus their effect of spreading on the volcanic pile is also removed). In some areas several volcanoclastic units as a group mark a hiatus in deposition of pillow lavas. Abrupt lateral change in volcanic facies is common within the formation.

Lithology

Most lavas are dark green to grey-green, aphanitic to fine grained, aphyric amphibolites. Some pillow lavas, however, are medium grained amphibolite with textures resembling those of doleritic dykes and massive flows. These

rocks may be spotted with hornblende crystals (such coarse-grained pillow lavas and dykes are present in columns 3 and 10, Fig. 8).

Vesicles and amygdulæ are rare in the Cameron River Basalt. They are well developed at one location, about half-way through the formation just east of column 5 (Fig. 8a). Here, smooth-sided, rounded to elongate vesicles, up to 20 mm in diameter (but generally less than 5 mm) occur in margins 40 to 130 mm wide around the pillows. The vesicular-rich margins are thicker along the top of the pillows. Coarse vesicles do not appear in the cores of pillows. Presumably fine and minute vesicles have been obliterated by the amphibolitization of the original basalt.

Some basaltic lavas weather very dark green to black and have an exceptionally smooth glacial polish. Rims of the pillows are extremely thin and difficult to recognize. These basalts, which are distinct from the common types, occur at several locations throughout the Cameron River Formation:

1. In the lower parts of the formation near the north end of Patterson Lake and west and southwest of Webb Lake; and
2. In the upper parts of the formation east of Allan Lake.

There is no distinct pattern to the distribution of this lithology.

The fresh rock is a very fine grained, dark green amphibolite, that has the highest specific gravity (3.04-3.05) and the highest normative plagioclase compositions (An₆₈₋₈₀) of all rocks in the Cameron River belt. Normative colour index is moderate to high. Amphibole, which makes up 75-90% of the rock, is ferroactinolite showing weak pleochroism in pale shades of blue-green (Z), olive (Y) and neutral (X), commonly with low birefringence (first order red to purple) and extinction angles (Z C of 15-19°). Ferroactinolite forms poikiloblastic, anhedral, equant to acicular grains (ranging up to 2 mm but generally averaging 0.5 mm) that enclose quartz, calcite and opaque minerals. Some rocks have two generations of actinolite: anhedral poikiloblasts (0.6 mm) with random orientation and elongate to ragged acicular grains (0.1-0.2 mm) that have preferred orientation. Epidote appears in trace amounts (ca. 0.2%) except in veins cutting the rock where it is abundant along with calcite. Felsic material between the amphibole is quartz and untwinned plagioclase. Rarely, relics of primary plagioclase are recognized by lath forms (0.1 by 0.5 mm) that are internally polygonized. No primary igneous minerals are present. Outcrop characteristics of this rock apparently result from the high amount of very fine grained amphibole.

Feldsparphyric units, both massive and pillowed, contain less than 1% phenocrysts of feldspar (3-10 mm) scattered randomly through the rock. The main lithological difference between the massive and pillowed units is the microscopic texture of the matrix. Massive units are coarser grained and contain relics of feldspar crystals defined by lath shaped patches (0.1 to 0.4 mm) of irregular feldspar grains that have similar extinction positions. Interlocking feldspar relics and hornblende crystals suggest an original

fine grained rock that had a diabasic texture. Blue-green hornblende forms coarse (0.4 to 2 mm) elongate crystals that contain only sparse inclusions. Presumably hornblende has replaced primary pyroxene. In contrast, pillowed units are much finer grained, plagioclase relics are not distinguishable amongst the irregular seriate granoblastic matrix and hornblende is crowded with inclusions of plagioclase, quartz and calcite to form a sieve texture.

Mafic volcanoclastic sediments. Mafic volcanoclastic rocks occur sporadically throughout the Cameron River Basalt, generally as thin (10-30 m) units that are not continuous laterally for any appreciable distance. The thickest and most extensive units occur midway through the formation about 4 km northeast of Dome Lake (columns 2 and 3, Fig. 8) and east of Allan Lake (columns 7 to 10, Fig. 8).

Metamorphic foliation develops more prominently in clastic units than in the adjacent pillow lavas. Locally, this foliation gives a weathered surface of clastic rocks an erroneous impression of being thin-bedded. Generally the coarseness of metamorphic recrystallization obliterates delicate primary textures.

Northeast of Dome Lake (columns 2 and 3, Fig. 8) two or three units of volcanoclastics, 15-16 m thick, occur at about the same stratigraphic horizon and are correlated laterally for about 1500 m. In some places they are intimately associated with the overlying pillow breccias. Outcrops vary from massive to sublayered in which preferred orientation of pitting on weathered surface, orientation of clasts, and in some places of metamorphic foliation, reflect a subtle layering. Clasts 1-2 mm across are common and some have relict textures resembling that of pumice. These rocks resemble lapilli tuff on weathered surface.

East of Allan Lake the upper part of the formation contains about 220 m of light grey, buff and green, vaguely layered, fine volcanoclastics and rare shale that appear to be a thick lens of clastic material which is not continuous laterally for more than about 1500 m. Thinner units near the top of the formation locally are well bedded. The lower half of the formation in this area contains 200-300 m of coarse- to fine-volcanoclastic sandstones. Although units vary from non bedded and poorly bedded to well bedded (0.5-1 cm), generally they are massive. One unit has crude bedding that grades upwards (within a few decametres) into a well bedded, fine clastic rock containing angular fragments of rhyolite. Invariably near the top and the bottom of this large lens the volcanoclastics are associated with the pillow breccias.

Thin, local units of volcanoclastics that are intimately associated with the pillow breccias are interpreted to be hyaloclastites. Thick successions, that are devoid of pillow lavas or pillow breccias, and comprise dominantly massive to vaguely bedded units with minor well bedded sequences may represent redeposited hyaloclastic material or epiclastic material derived from the immediate volcanic pile. The massive nature of the volcanoclastic units is compatible with deposition by turbidity currents or mass flows. Rhyolite conglomerate and rhyolite fragment wackes enclosed within the mafic volcanic succession are evidence that at least some material was not locally derived.

Shale and siltstone. Lentils of shale and volcanic sediments are intercalated with mafic pillow lavas for a stratigraphic thickness of about 130 m near the contact between the Cameron River Basalt and the Yellowknife sediments, about 500 m from the southern end of Fenton Lake. East of Allan Lake several lenses of black shale or slate that occur at irregular intervals within the Cameron River Basalt, contain pyrite (usually expressed as small gossan zones on weathered outcrop) and possibly graphite.

The thickest and most persistent units of dark greenish grey to dark grey siltstone and shale within the Cameron River Basalt form a series of lenses traceable for about 5 km along a horizon midway through the volcanic belt and ending near Bambi Lake. The unit pinches and swells along strike, varying from 10 to 50 m thick. In many places, poor outcrop prohibits thickness determinations and there may be more than one unit. This unit is at about the same horizon as lenses of rhyolite boulder conglomerate and rhyolite-bearing wacke that appear at both ends of the shale. Volcanic rocks enclosing the siltstone-shale succession are generally pillow lavas and pillow breccias and coarse volcanoclastic sediments associated with them.

In one place near the western end of the unit, a 51 m section comprises a 10 m thick basal unit of pale to medium grey cherty sediment; 7 m of rusty weathering, pyritiferous, black slate that makes sharp contact with the underlying chert; and 34 m of thin-bedded, dark grey shale and siltstone that contain pyrite and magnetite. Pyrite- and magnetite-bearing shales commonly weather rusty brown in patches. Coarser beds in the upper siltstone units have scour channels and graded bedding. The shale-siltstone successions contain interlaminated (0.5-5 mm layers) dark grey shale and poorly sorted, medium grey siltstone, that make up 38 and 62 % of the rock respectively. The heavy and mafic mineral content of the whole rock is about 7 % amphibole, 15 % biotite, 8 % chlorite, 4 % pyrite and 1 % magnetite. Blue-green pleochroic hornblende forms randomly oriented, poikiloblastic anhedral (0.4-1.5 mm) with rims of colourless tremolite that have grown across the bedding. This mineral is most abundant in shale layers (15 %) and is sparse in silty layers (2 %). Anhedral patches (20-50 microns across) of brown biotite that have random orientation are also more abundant in the shale layers (20 %) than in silt layers (12 %). Pyrite occurs in both layers in equal amounts as lenses 0.2-1.0 mm long that lie in the bedding plane. Magnetite forms closely spaced clusters and trains of minute dust-like grains that make up about 3 % of the shale layers and 1.5 % of the silty layers.

In one place near the centre (of strike length) of the unit, pelitic beds are deformed into small open folds of amplitude about 2.5 m. This unit lies between deformed, coarse, volcanoclastic rocks (possibly hyaloclastites or related pillow breccias) in which layering is defined by thin (2-10 cm) undulatory, pinching and swelling lenticles that warp around larger lensoid clasts. The pelite is a 10-12 m thin-bedded unit consisting of alternating biotite-hornblende-rich and quartz-rich siltstone laminations (2-10 cm thick) that display contorted undulatory layers. Crests of minor folds form a prominent lineation that plunges 85° towards 335°. Microscopic crenulations (with amplitude 0.2 mm and

wavelength 0.5-1 mm) in dark layers are outlined by biotite along layering and strong preferred orientation of biotite parallel to the axial surfaces of the crenulations. In general, the finest shale layers are rich in biotite and hornblende (35-40%) whereas silty layers contain dominantly hornblende and minor biotite and the total mafic content is less than about 5-10%.

One outcrop near the eastern end of the unit contains 20 m of interbedded greywacke, siltstone and shale that form beds 3-60 cm thick. Near the base of this succession some 1-3 cm thick beds, of carbonate-bearing arenite contain pebbles of quartz up to 1 cm across.

In several places along this horizon the shale-siltstone succession superficially resembles iron-formation because of its dark colour and apparent high specific gravity. Apart from the fact that locally it contains appreciable amounts of amphibole and biotite, mineralogically it is not considered to be iron-formation.

Rhyolite fragment conglomerate and volcanics. A lens of rhyolite boulder conglomerate, about 85 m thick and 1100 m long, lies midway through the formation between Milt and Allan lakes (almost at the same horizon as the shale-siltstone units). Near the western end, the lens is a very poorly sorted, massive conglomerate comprising 10-25% blocks and rounded to subrounded boulders and pebbles (ranging from 1 to 60 cm) in a dark green-grey amphibolitic matrix containing fine clasts of rhyolite and minor quartz. In general, however, large blocks and boulders make up less than 5% of the unit, and most of the clasts are less than 5-10 cm. Most clasts are elongate to give the outcrop a steeply plunging (ca. 60°) lineation. Near the thickest part, the lens is a polymictic conglomerate that contains subrounded to rounded pebbles and boulders up to 25 cm across (but generally 0.5-2 cm) that are 60-70% metabasalt (very fine grained amphibolite), 20% aphanitic rhyolite and rhyolite porphyry (containing quartz and feldspar phenocrysts to 0.5 cm). In one place the lower part of the unit grades eastward into arenite and finally into black siltstone and shale. The finer parts of this unit look more like a pyroclastic rock than a conglomerate, and possibly could be a subaqueous mass flow or lahatic deposit. In at least one place the matrix of this boulder-bearing, massive volcanoclastic unit contains rhyolite clasts (up to 1 by 3 cm) in strong preferred orientation having very irregular boundaries and irregular terminations that mould around tiny lithic clasts. The texture of this rock is virtually identical to the eutaxitic foliation seen in recent deposits of welded tuffs. In view of the position of this unit within a thick succession of pillow lavas that were undoubtedly deposited in a subaqueous environment, it may be rationalized that this is indeed a tuff or a volcanoclastic rock, but the fiamme-like features were formed as a result of tectonic flattening of pumiceous material rather than by plastic deformation of hot pyroclastic material.

A unit of rhyolite-fragment volcanoclastic rock, 2-10 m thick and about 2000 m long, lies between mafic pillow lavas and pillow breccias about 30 m above the rhyolite boulder conglomerate unit. The unit is similar to the matrix of the rhyolite conglomerate. This pale grey to pale green

locally brownish-grey weathering unit weathers rusty brown along its margin where it contains pyrite. It is a poorly sorted volcanoclastic rock that lacks internal layering. The rock contains about 30-50% buff, pale and medium grey weathering lensoid fragments up to 20 by 5 mm, but generally less than 5 by 1-2 mm, that have strong preferred orientation in a dark green-grey hornblende-rich matrix. Composition of the clasts is essentially the same throughout the unit and all are recrystallized relics of glassy or crystalline volcanic material. About 75-80% of the fragments are rhyolite and dacite, 20-25% are andesite, and 5-10% are quartz. Andesitic and some dacitic fragments are subangular to subrounded and commonly elongate, whereas rhyolitic fragments vary from lensoid to oval, and some have irregular and cuspidate boundaries resembling shards and pumiceous material. In thin section the rock is almost completely recrystallized and the original compositions of fragments are inferred from the crystallization textures, relative proportions of metamorphic mineral species and their comparison with rocks of known chemical composition. Meta rhyolitic clasts comprise very fine (0.01-0.05 mm) microcrystalline aggregates of quartz and feldspar and less than 1% amphibole. These felsitic fragments are interpreted as recrystallized, devitrified shards of glass. Hornblende-poor (present as 5-10% radiating clusters) fragments that contain thin lath-shaped plagioclase phenocrysts in microfelsic matrix may be of dacitic composition. Andesitic fragments contain euhedral, rectangular to blocky microphenocrysts of oligoclase up to 0.8 mm in a microfelsic matrix that has randomly oriented relics of plagioclase microlites. Anhedral grains of blue-green hornblende with inclusions of minute round felsic grains and fine radiating splay of amphibole, make up about 25% of the fragments. All the amphiboles are considered to be the result of metamorphic crystallization. Although plagioclase microlites and some phenocrysts are polygonized and recrystallized, domains of similar extinction mark the crudely rectangular or lath forms of the original crystals. Large (about 0.7 mm) microphenocrysts commonly have minutely irregular boundaries, but their cores are not polygonized. Quartz forms rounded diamond shapes and rounded to subrounded grains with undulose extinction or coarse polygonized aggregates. These areas are possibly recrystallized quartz phenocrysts. The matrix to the fragments is a hornblende-rich mixture of hornblende and microcrystalline quartz and feldspar that has no distinct texture. The matrix may be recrystallized, very fine clastic material of basaltic composition.

Most of the irregular, cuspidate microfelsic rhyolite clasts are considered to be recrystallized, devitrified glass shards of pyroclastic or hydroclastic origin, whereas rounded to subangular andesitic clasts are considered to be lithic fragments of epiclastic origin. Poor sorting and lack of layering in the unit over a distance of 2 km is compatible with deposition as a mass flow. The stratigraphic position of this unit within a very thick succession of pillow lavas suggests that the clastic material was deposited in a subaqueous environment.

This unit is interpreted as a subaqueous mass flow, possibly initiated by explosive eruption of rhyolitic magma into the sea. The turbidity current comprised a suspension of

shards, pumice, lapilli of rhyolite and andesite, and epiclastic volcanic material derived from unconsolidated volcanic deposits near or at the site of the explosive eruption. The rhyolite boulder conglomerate lens, commonly only 30 m below this unit, is evidence that a rhyolite body, possibly a subaqueous dome, existed in this vicinity even though it is not exposed in the present section. Strong preferred orientation of particles may be a result of tectonic flattening or stretching, which is clearly evident in the enclosing pillow lavas.

Several other units of massive rhyolite and rhyolite volcanics that occur in the lower part of the Cameron River formation are isolated from any of the main bodies of rhyolite. Rhyolite volcanoclastic units are almost invariably associated with mafic volcanoclastics in the pillow lava succession. These white to pale grey rhyolitic to dacitic beds, some containing quartz and feldsparphyric rhyolite and opalescent quartz eyes, generally less than 3 to 5 m thick, occur within a thick unit of mafic volcanoclastic rocks 500 to 700 m below the rhyolite boulder conglomerate, between Milt and Allan lakes. One 35 m thick massive unit grades upward into well bedded tuffaceous-looking rock containing angular rhyolite blocks.

Near the west-central side of Webb Lake, one unit of massive rhyolite porphyry appears to be a dyke rather than a lava flow or tuff. A large gabbro dyke has intruded the rhyolite dyke, with the effect of splitting it apart longitudinally into two slivers. The thickness of the two slivers combined is about 40 m. The rhyolite is intensely sheared locally to form a well layered mylonite. The rhyolite comprises phenocrysts, as much as 3 mm across, of rounded to embayed quartz (5%) and euhedral to embayed albite (An₇) (25%) in a microfelsic matrix containing tabular microphenocrysts of plagioclase up to 0.3 mm long. The matrix has a distinct fluidal texture defined by discontinuous biotite-rich filaments and lenticles and microphenocrysts that stream around the larger phenocrysts. Relict phenocrysts of quartz and plagioclase have well developed pressure shadows, defined by recrystallization products slightly coarser than the matrix, that lens out parallel to the foliation. A sheared variety of this rock has fine (1 cm wide) layers with sharp boundaries and a pronounced mylonitic foliation.

Interpretation

The Cameron River Basalt represents voluminous eruption of basaltic magma in a subaqueous environment. The dense swarm of amphibolite dykes and sills, some lithologically identical to pillowed lavas (eg. feldsparphyric units), that permeates the volcanic pile but not the overlying Burwash Formation, is assumed to be part of the plumbing system along which the lava erupted.

Although pillows face westward in almost all cases, it is not clear that this homoclinal succession was ever a vertically stacked pile of lavas. Perhaps the present erosion surface exposes a view of the volcanic pile perpendicular to the general flow direction so that we now see a deformed agglomeration of originally overlapping lavas that accumulated successively outward from east to west.

It is of interest to speculate on the paleogeographic submarine setting which caused the mafic volcanoclastic deposits to form thick localized pods with restricted lateral distribution in parts of the volcanic belt. Were there basins or channels in the pillow lava succession into which sediments were deposited? One model is that the sediments were deposited within the volcanic pile in submarine canyons transverse to the volcanic belt. During volcanic eruption, earthquakes may have triggered turbidity currents and submarine avalanches. Possibly these avalanches eroded unstable portions of the lava pile along penecontemporaneous fractures, created by the earthquakes, to form submarine channels. Once formed, such canyons may have remained the locus of submarine erosion and deposition as the lava continued to accumulate.

Local units of shale and siltstone record short pauses in deposition of lavas. Clastic units and dykes of rhyolite within the formation suggest that small amounts of rhyolitic material erupted between the major effusions of basaltic magma. The source of most of these rhyolite volcanoclastics is not exposed.

Webb Lake Andesite

Definition, distribution and thickness

The Webb Lake Andesite comprises pillow lavas and pillow breccias of metamorphosed andesite and basaltic andesite and associated volcanic arenites and rudites. Local lenses of rhyolite breccia near the western margin of the belt are included in this formation.

West and southwest of Webb Lake the formation apparently interfingers with the Cameron River Basalt across a width of 1400 m, thins to about 400 m between columns 2 and 5 (Fig. 8) and ends about 1100 m south of column 2. The thickness is variable near the southern end where the formation overlies rhyolite domes. In places metagabbroic sills and dykes have "spread apart" portions of the formation so that locally the stratigraphic thickness may be much less than that shown in Figure 8a. For example, the lowest lobe of the Webb Lake Andesite east of column 6 is as much as 100 m thinner than shown.

Along the eastern side of Allan Lake the formation pinches and swells and has a maximum thickness of 300 m. Although the northern extent of the andesitic unit is not known with certainty, it appears to taper out into well layered "greenstones" that lie at the top of the volcanic belt above sheared pillow breccias and associated clastic rocks. The unit may continue farther to the north than shown on the map.

Contact relations

Contacts between the Webb Lake Andesite and the Cameron River Basalt generally are conformable. In places the contact is marked only by an abrupt change in lithology and there is no change in the structure of the pillowed units. Locally a gossan zone (containing abundant, finely disseminated pyrite) clearly marks the upper margin of andesite fingers in the Cameron River Basalt. In other places, along

the eastern side of the formation east of Allan Lake, the contact occurs at a break in slope. Near column 5 (Fig. 8a) andesite pillow lavas and pillow breccias form a ridge that is marked on the east side by a prominent north trending valley.

Lenses and layers of massive rhyolite and rhyolite breccia are common along the eastern side of the Webb Lake Andesite. The boundary between rhyolite-bearing, andesite fragment breccia and underlying rhyolite breccias, between columns 4 and 5 (Fig. 8), appears to be conformable although the contact has not been defined precisely. In places this contact appears gradational. Near column 9 (Fig. 8), lenses of rhyolite fragment breccia or conglomerate mark the contact between the Cameron River and Webb Lake formations. This breccia contains clots, lenses and irregular streaks of rhyolite (10-40 cm wide and up to 1 or 2 m long) in a mafic matrix.

The lower contact of the Webb Lake Andesite is displaced by a series of normal faults with down-dropped sides to the north between columns 5 and 6 (Fig. 8). Except for one fault these displacements do not appear to be reflected in the upper (western) contact of the formation.

The contact between the Webb Lake and the Burwash formations generally follows Cameron River between Allan Lake and Dome Lake. It is exposed at many places near the base of bluffs along the east side of the river, as well as on the protuberance in the volcanic belt about 7 km southwest of Allan Lake and at the southwest end of Allan Lake. In all places the contact appears conformable or volcanic breccias interlayer with siltstones and wackes.

A gossan zone, 2-10 m wide, within the adjacent sediments commonly highlights this contact. This pyrite-rich gossan weathers variably orange, deep red brown, or maroon. The darker colours appear where units of mafic volcanic breccia interlayer with the sediments. Locally, sediments in the gossan are silicified.

Exact contact between andesites and metasediments generally is not exposed along Allan Lake. Near of the contact, however, the trend of bedding in metasediments and in the clastic volcanics are parallel, suggesting that the units are conformable. Andesite near this contact changes from dominantly pillow lavas to pillow breccias, hyaloclastites, and volcanic pebble conglomerate. The contact is exposed on a small point at the southwest end of Allan Lake where it drains into Cameron River. At this locality folded schistose sediments make sharp conformable contact with flattened and stretched andesitic pillow lavas. Steeply overturned bedding and pillow lavas face westward, (7 km southwest of Allan Lake).

Along the north side of the protuberance a narrow depression marks the contact between volcanics and sediments. The completely conformable nature of the contact is distinguishable in spite of the folded nature of the rocks.

In places the contact is a gradational transition from rhyolite fragment breccia through pebbly grit to wackes. This transition is exposed on an islet in Cameron River about 6 km from the south end of Allan Lake. Here, breccia fragments and pebbles are quartz-phyric rhyolite similar to

that in the rhyolite breccia near the top of column 5 (Fig. 8). The breccia at the transition comprises subrounded to subangular clasts of pale grey rhyolite (up to 20 cm across, but generally less than 5 cm) and lesser amounts of bedded siltstone in a rhyolite pebble grit matrix. Some fragments have abundant cavities in the centre, but massive rims. One large fragment that has undulose boundaries and a cavity-rich centre could be a volcanic bomb. Units are massive and show no clear bedding, except for a gradation from cobble breccias to pebbly siltstones and wackes. Although clasts appear equant on the surface, on vertical faces they are elongated and form a steeply plunging lineation, indicating considerable deformation.

Contacts between andesites and mafic intrusions (metadolerite or metagabbro dykes and sills) are sharp. Some intrusive margins contain schlieren of andesite.

Internal structures

Most of the formation comprises thick successions of pillow lavas and pillow breccias that are devoid of layering. Layering can be determined in a crude way by tracing gradational contacts between pillow lavas and pillow breccias and rarely where thin beds of arenaceous clastic material occur within the succession. Near column 7 (Fig. 8) layering is locally defined by alternating thin lobes of massive lava and breccia. Bedding in the arenaceous clastic rocks (probably hyaloclastites) is not well defined or is inferred from the preferred orientation of clasts, which may be an erroneous inference in the light of the intense deformation. Crude layering, moreover, may have been rendered unrecognizable by the intense deformation. Thin-bedded volcanoclastic rocks are rare. Such contacts and beds generally are steeply dipping and parallel to the trend of the volcanic belt and to the direction of flattening in the pillow lavas. Between columns 2 and 4 (Fig. 8) attitude of flattening and layering is steeply dipping and pillows face easterly, which is the general case in this part of the volcanic belt, in contrast to westerly facing units in the central and northern parts of the belt.

In general, pillow lavas are closely packed, sack-like bodies (in two dimensions) that mould around each other and have little or no clastic material between them. In most places they are flattened and stretched. Pillow breccias at the top of the formation at the south end of Allan Lake contain fragments that are extremely elongate but only slightly flattened. In cross-section the fragments appear undeformed and are equant, subrounded to subangular (rarely angular) and average 3-10 cm across, with the occasional block up to 20 cm across. Some fragments have curved lunar shapes or are rounded with curved tapering terminations. The fragments are drawn out into rods that have length to breadth ratios of about 5:1. Near Allan Lake the steeply dipping (about 75°) pillow lavas and pillow breccias are overturned to the west.

The steep plunge of the elongate clasts is regular and consistent over a broad area, suggesting that the elongation is tectonic. In places clasts are both intensely flattened and stretched. Commonly, a weak schistosity and in places a slaty cleavage develop parallel to the trend of flattening of

pillows. Some outcrops weather with deep pock marks that outline the trend of flattened pillows and schistosity. The flattened pillows have constant preferred orientation over broad outcrop areas.

In one area the intersection of three steep planar features causes outcrops to fracture into almost vertical pencil-like slivers. These planar features are: (1) direction of pillow flattening and stretching, (2) strongly developed schistosity (S_1) almost parallel to the direction of pillow flattening, and (3) a second cleavage (S_2) comprising weak, discontinuous fractures that are axial planar to small folds that deform S_1 .

Folding on a small scale is evident in parts of the formation (see Structural geology). About halfway between columns 5 and 6 (Fig. 8) at the top of the formation, both pillows and cleavage parallel to flattening are folded into a small open fold of amplitude about 2 m. There does not appear to be any major repetition of units by folding. In Figure 2 structural symbols on pillow lava successions indicate the direction of facing of the pillows as well as the plane of flattening of the pillows. The plane of flattening, because of its constant orientation over broad areas and parallelism to schistosity, cleavage and stretching of pillows, is a tectonic feature and may not reflect primary attitudes in the pillow succession.

Stratigraphy

Northeast of column 6 (Fig. 8) lobes of andesite that interfinger with the Cameron River Basalt are pillow lavas except for the top of the section where the andesite is a massive unit, and in the lowest tongue northeast of the section where the upper 35-50 m is massive to thin-layered volcanoclastic arenites that cap a 120 m succession of pillow lavas. The volcanoclastic rocks are interpreted as meta-hyaloclastites.

Between columns 5 and 6 (Fig. 8) the bulk of the formation is dominantly close-packed pillow lavas with lenses of pillow breccia and minor layered clastic units. Near column 5 (Fig. 8) the lower half of the formation is pillow lavas that pass upwards into several cycles containing alternating closely packed pillows and intervening pillow breccias. In places the pillow breccia forms the upper part of the cycle; in others it is the lower part. Each two-part cycle is about 150-200 m thick. Three cycles can be distinguished in column 5 as follows. The lower cycle has a base of breccia (about 160 m thick) that grades upward through pillow breccia to closely packed pillow lavas which make up most of the unit (approximately 200 m). The top of the cycle is marked by a distinctive 13 m thick bed of pillow breccias in which pale greenish-grey weathering stretched pillows and blocks of andesite lie in the brownish-grey weathering fine clastic matrix (hyaloclastite). The middle cycle in this succession is interbedded pillow breccias and pillow lavas in about equal amounts, whose top is marked by coarse volcanic arenite. Contacts are gradational and no discrete beds are recognizable. Layering, however, is suggested by the preferred orientation of fine fragments and the flow structure in the matrix around them. The uppermost cycle com-

prises basal pillow lavas that grade upward into closely packed pillow lavas.

Two protuberances between columns 4 and 6 (Fig. 8) extend 200-700 m west of the gently undulating western margin of the Cameron River belt and represent local stratigraphic piles up to 500 m thick. The northernmost protuberance contains highly deformed, foliated to schistose amphibolite that varies from thin laminated to lenticular (deformed breccias?) and possibly pillowed rocks. These rocks appear to be more mafic than typical andesites of the Webb Lake formation. Column 5 (Fig. 8) shows the stratigraphic succession through the southern protuberance constructed in part from field observations provided by W.R.A. Baragar (unpublished information). Baragar's data show a basal 80 m thick unit of basaltic andesite pillow lavas overlain by 140 m of rhyolite breccias that are capped by 30 m of andesitic breccia. The rhyolite breccia forms a lenticular body about 500 m long that is convex westward in rough conformity with the western side of the formation. The rhyolite fragment breccia gives way to andesite fragment breccia along the southwestern side of the lens but appears to continue up to the contact with Burwash sediments at the northwestern side.

Between columns 3 and 5 (Fig. 8) the formation is a mixture of pillow lavas with lenses of andesitic arenites and rudites, rhyolite fragment breccias, and massive units of quartz-phyric andesite. The internal stratigraphy in this part of the formation has many abrupt lateral changes so that the proportions of pillows to pillow breccia at a particular horizon in one section may be quite different from the proportions in a section only 200 m away. Between columns 4 and 5 (Fig. 8) the lower half of the Webb Lake formation is dominantly felsic fragment breccias, and the upper half is pillow breccias and hyaloclastite. The lower breccias are poorly sorted and unlayered and contain pale grey weathering fragments, most of which range from 2 to 60 cm (W.R.A. Baragar, unpublished information). Some fragments contain abundant amygdules that have fillings of chlorite and calcite. In outcrop the fragments contrast with the dark grey to green fine clastic matrix.

The Webb Lake Andesite, above the rhyolite domes southwest of Webb Lake, is mainly aphanitic grey andesite pillow lavas. In column 3 (Fig. 8), however, the lower 200 m is massive quartz-bearing andesites and the upper 110 m contains arenaceous volcanoclastic rocks. The volcanoclastics locally contain fragments of quartz 2-3 mm across with some pebbles up to 2 cm across (W.R.A. Baragar, unpublished information). In one place the unit is thin-bedded, but in general bedding is not well defined. The apparent vague bedding in the clastic rocks may be a manifestation of deformation rather than a primary feature. Conversely, deformation may have rendered any delicate bedding features unrecognizable. Flattened and stretched pillow lavas occur between the massive and clastic units.

The lobe of the formation north of column 6 (Fig. 8) is dominantly pillow lavas that contain lenses of pillow breccia, hyaloclastite and rudaceous volcanic arenites between columns 9 and 10, and a poorly defined upper zone of volcanoclastic rocks along the western side of the unit near

columns 7 and 8. The lenses of pillow breccia and hyaloclastites are 20 to 40 m thick and occur within closely packed pillow lavas. The unit grades over 3 m from pillow lavas to pillow breccias to rudaceous arenites.

Near column 7 (southeast end of Allan Lake) the formation comprises 110 m of closely packed pillow lavas capped by 60 m of pillow breccia. The unit changes gradually from a mass of closely packed pillows to several irregular masses and tongues of pillow lava, with intervening pillow breccias, to dominantly pillow breccias with rare pillowed flow lobes (approximately 2 m thick).

The proportion of clastic material increases northward, such that where the formation thins to 40 or 50 m between columns 8 and 9 (Fig. 8), it consist almost entirely of volcanoclastic rocks.

Lithology

In comparison to metabasaltic rocks, meta-andesites are lighter coloured on both weathered and fresh surface, finer grained (mostly aphanitic), contain less amphibole (generally finer grained but lighter coloured than in the basalts), and have a lower range of specific gravity (2.70-2.80). The main lithological types include (1) aphanitic pale grey to medium grey weathering massive or pillowed units and associated pillow breccias and hyaloclastites that have mafic selvages, (2) quartz-phyric andesites, (3) sparsely porphyritic, felsic (or leucocratic) meta-andesites, (4) felsic andesite breccias and arenites, and (5) mafic andesites.

Most meta-andesites are pale- to medium-grey or greenish-grey weathering aphanitic rocks that are slightly darker (medium- to dark-grey or greenish grey) on fresh surface. Some andesite pillow lavas have dark green rims, 5-15 mm thick, containing about 40 % medium grained (1-4 mm long) blue-green hornblende.

The rocks are commonly microporphyritic, containing 1 to 2 % phenocrysts (0.2-0.5 mm across) of plagioclase that form euhedral to subhedral slender laths and stubby, blocky crystals. Rarely, they form glomeroporphyritic clusters. Some phenocrysts have rounded to partly embayed outlines. Subhedral and some anhedral phenocrysts have minutely irregular boundaries that are intergrown with microfelsic matrix. Primary polysynthetically twinned phenocrysts have fine turbid alteration in contrast to recrystallized plagioclase in the granoblastic matrix which is clear. Most microphenocrysts have even extinction, although some are partly polygonized (recrystallized). The most calcic phenocrysts are andesine that have weak normal zoning (An_{34-48}). Most, however, An_{34-38} are albite (about An_8). Feldspars of this composition probably formed during metamorphism and do not represent a primary igneous composition.

Andesine microlites (An_{37}), showing little polygonization may be original igneous feldspars. Most of the microfelsic matrix in these rocks, however, consists of highly polygonized relics of microlites. Although microlites are recrystallized, twins are still recognizable. The relics have crude lath forms (ranging from 0.01-0.3 mm) with irregular boundaries, comprising minute (1-40 microns) subgrains of albite that have

one dominant extinction position within each twin unit of the original crystal. The slight variation in grain size in polygonized microlites, the irregular boundaries between some crystals and the amoeboid areas within partly polygonized crystals form an overall felsitic aggregate that has an irregular granoblastic texture. Felted to trachytic fabrics are still recognizable within this recrystallized mass.

Amphibole, which makes up from 15 to 30 % of meta-andesites, appears to be entirely of metamorphic origin. Its habit varies from anhedral, poikiloblastic, elongate patches to radiating sheaves of acicular crystals. The most common amphibole is blue green (pleochroic scheme $X = \text{neutral}$, $Y = \text{olive green}$, $Z = \text{blue green}$; $ZC = 12-13^\circ$). Poikiloblastic hornblende has grown across the included parts of the microfelsic matrix material as well as opaques, biotite and epidote. Many rocks show preferred orientation of elongate grains or clusters of grains. Colour of the rocks seems to vary with the species of amphibole, its amount and size. Pale grey meta-andesites tend to have very pale green hornblende, or almost colourless tremolite, whereas the darker rocks have coarser grained medium blue-green amphibole in amounts greater than 25 %.

Biotite makes up generally less than 1 to 2 % but is up to 8 % in some rocks. Some of the biotite and invariably chlorite is a secondary alteration of amphibole. In other cases biotite is a metamorphic phase that imposes schistosity to the rocks.

Accessory minerals include epidote, zircon, apatite, and opaques. Opaque minerals are most commonly magnetite but also include hematite, ilmenite and pyrite.

The most common metamorphic mineral assemblages are: quartz-albite-hornblende-biotite-magnetite (ilmenite) and quartz-albite-hornblende(tremolite)-biotite-epidote-magnetite. Most rocks contain some chlorite and calcite.

Felsic andesites. Felsic meta-andesites are buff to very pale greenish-grey weathering rocks that are aphanitic or sparsely microporphyritic. These light- coloured rocks appear to be much more siliceous than other meta-andesites. In thin section they are similar to the common variety of andesite except that the amphibole, which makes up 15 to 20 % of the rock, is very pale green actinolite instead of hornblende. Actinolite typically forms slender to stubby needles or radiating sheaves of needles less than 0.5 mm long. The matrix is an irregular granoblastic microfelsic aggregate in which polygonized relics of feldspar microlites are identifiable.

Some of the pillow breccias and breccias within the upper 200 m of the formation between columns 4 and 6 (Fig. 8) contain pale greenish-grey weathering fragments (that are medium grey, very fine grained or aphanitic rocks on fresh surface), that appear to be more felsic than the surrounding meta-andesites. Based on field study, some of these rocks were thought to be rhyolitic or dacitic breccias. In thin section, however, fragments of these leucocratic breccias are thoroughly clouded with intense turbid alteration comprising minute (2-5 microns) grains of epidote and very pale greenish-neutral laths and scales possibly of amphibole or micaceous clay minerals. Some of the rocks also contain about

15 % epidote as idiomorphic patches or poikiloblastic grains (up to 0.5 mm across) and 5 to 10 % very fine calcite evenly scattered throughout. Some fragments contain about 30 % blue-green hornblende but the turbid alteration overprints all of the minerals and accounts for the felsic appearance in hand specimen. Some of the breccias are sheared so that the felsic lenses comprise minute microcrystalline streaks surrounded by the turbid alteration. Amphibole is generally lacking in such rocks, whereas chlorite and epidote (and in one case phlogopite and magnetite) are abundant.

A breccia at the constriction in this formation between columns 8 and 9 (Fig. 8a) contains 40 % felsic fragments in a clastic matrix that has about twice as much hornblende as the fragments. The rounded to subangular fragments contain up to 50 % lenses of granoblastic quartz interpreted as recrystallized amygdulites (but could be relict phenocrysts). These fragments may be metadacites or amygdaloidal felsic andesites. Breccias containing similar clasts occur within the upper lens of the fragmental rocks in column 9 (Fig. 8).

Quartz-phyric andesites. Quartz-bearing andesites form massive units and fine grained clastic unit rocks near the top of columns 3 and 6 (Fig. 8). In column 3, the lower massive andesite unit (which may be several depositional units), increases in quartz content upwards. The upper part of the formation in this section is crudely layered fine clastic rocks containing pebbles up to 2 cm across. These pale green, buff to rusty weathering rocks contain 2 to 10 % opalescent blue-grey eyes of quartz and 1 to 3 % altered phenocrysts of feldspar in a dark grey aphanitic matrix. Quartz phenocrysts are rounded, ovoid to blocky forms, ranging from 0.2-2 mm (but occasionally up to 5 mm) that have minutely irregular boundaries intergrown with the microfelsic matrix. Most are single crystals but some are polygonized into three to five subgrains. In one of these rocks plagioclase phenocrysts (0.5-1.5 mm) are rounded, rectangular to kidney-shaped forms that appear to be partly resorbed crystals. They typically have fine sericite and calcite alteration and some have a rim of submicroscopic turbid alteration, suggesting zoning. The lavas show relict flow textures (relics of microlites that form felted fabrics or stream around the plagioclase phenocrysts) whereas clastic rocks have a felsic microcrystalline matrix where no relics of plagioclase are recognizable. Perhaps the clastic rocks comprise recrystallized glassy material and are thus hyaloclastites or possible hyalotuffs. Amphibole is lacking in these quartz-phyric rocks, but they contain 5 to 15 % phlogopite or biotite along with fine calcite in amounts up to 20 %.

Mafic andesites. Some of the pillow lavas and clastic rocks near column 4 (Fig. 8) are more mafic than the common type of meta-andesites. These rocks, which contain 40 to 55 % amphibole and have specific gravities ranging between 2.86 and 2.94, are probably metabasalts.

Rhyolite breccia. Dark grey to green-grey bluffs of rhyolite breccia at the top of column 5 (Fig. 8) comprise rhyolite fragments in a matrix of dark green amphibolite. Fragments are abundantly porphyritic containing up to 40 % quartz

phenocrysts in a felsic matrix having only minor mafic constituents.

Rhyolite clasts are not evenly distributed in the unit. Their abundance varies from 20 to 50 % of the rock and generally decreases westward. Locally they tend to be concentrated in layers (one such layer is 0.3 m wide) and the intervening material is fine mafic volcanic clastic rock carrying few fragments.

Clasts are equant to slightly elongate (1-10 cm; generally less than 5 cm) on horizontal surfaces but highly elongate (35-40 cm) in vertical exposures. These tectonically stretched clasts plunge vertically.

Interpretation

The Webb Lake Andesite formed entirely in a subaqueous environment, as indicated by pillow lavas throughout the succession and the position of the formation between basaltic pillow lavas of the Cameron River Basalt below and grey-wacke turbidites of the Burwash Formation above. Although minor unconformities are to be expected in a complex succession of this nature, no major unconformities were recognized. The clastic units in the Webb Lake Formation are believed to be pillow breccias and related hyaloclastites. Probably some are submarine debris flows. No unequivocal pyroclastic rocks were identified. Rhyolitic components in the breccias near the base and locally near the top of the formation indicate that debris was derived from both rhyolite domes and andesite lavas, that possibly accumulated while the pile of lavas was building up or covering the domes. Some rhyolite fragments may simply have been incorporated into the lava as it erupted through the rhyolite complexes. Interfingering of the Webb Lake Andesite and the Cameron River Basalt over a considerable distance may suggest that a major centre of andesitic volcanism was erupting near the southern end of the volcanic belt while major basaltic effusions were continuing to accumulate farther north. Alternately, and perhaps more likely, the apparent interfingering reflects the first flows of the Webb Lake Andesite that have filled in small valleys in the irregular topography of the underlying Cameron River Basalt (see Fig. 52c). Flattening of the belt along steep planes produced the interdigitating pattern seen in the present erosion surface. The association of rhyolite with the andesites might indicate that a differentiated magma chamber was being tapped, in which rhyolitic differentiates were the first to erupt, followed by more mafic andesites and basaltic andesites that formed parts of the prominent highs in the succession between columns 4 and 6 (Fig. 8).

Rhyolite

Three lenticular bodies of rhyolite, interpreted as domes and associated flanking breccias, lie at or near the top of the Cameron River Basalt northeast of Dome Lake (columns 1 to 5, Fig. 8) and south of Fenton Lake (between columns 10 and 11).

Dome south of Fenton Lake

This rhyolite forms a lens tapering from 400 m near the south end, to 30 m wide about 200 m to the north, then continues northeastward as a 10 to 30 m thick band for at least another 2000 m. The western side is gently undulating, whereas the eastern side is strongly convex. Where the body tapers out it is conformable in trend to the direction of flattening in the highly deformed pillow lavas to the west of it and to the coarse grained metagabbro sill on its east side. The unit in these areas is invariably sheared and weakly schistose. Steeply dipping schistosity is parallel to the trend of the unit.

In its thickest part it is buff to white weathering, medium grey, porphyritic rhyolite containing 10 to 20 % phenocrysts of quartz and 5 to 10 % phenocrysts (up to 1.5 mm across) of albite and oligoclase (An_{2-15} in a microfelsic matrix. Quartz phenocrysts are euhedral to deeply embayed rounded crystals 0.2 to 2 mm across. Some are polygonized so that a euhedral outline encompasses a granoblastic mass of subgrains. Euhedral to subhedral phenocrysts of albite and oligoclase (An_{2-15}) commonly have two-unit polysynthetic twinning and are not zoned. Although some plagioclase phenocrysts appear fresh, generally they have turbid alteration or are replaced by sericite and unidentified clay minerals. Matrix is an irregularly intergrown, granoblastic mosaic dominantly of quartz and untwinned feldspar having crystal sizes from 2 to 50 microns (most commonly about 20 microns) and minor amounts of sericite, biotite, chlorite, calcite, and trace hematite, magnetite, and zircon. Altered varieties contain up to 20 % sericite.

In the area where the main body tapers abruptly it comprises 60 m of sheared, rhyolite underlain by 50 m of quartz- and rhyolite-fragment breccia. Characteristic of this sheared rhyolite are the bright opalescent blue-grey quartz phenocrysts and an aphanitic matrix that varies from a buff, pale green to pale yellowish-grey rock having a greasy luster. Locally finely sheared and schistose zones weather in recession to give the outcrop the appearance of being crudely layered.

Breccia that underlies the rhyolite contains elongate fragments of sheared rhyolite and granitoid material, up to 10 cm across and 50 cm long, 15 to 20 % rounded quartz grains (resembling phenocrysts in the underlying rhyolite), 1 % feldspar crystals (to 1 mm) and pink garnet. Some granitoid fragments are well rounded. The matrix is a schistose (biotite, muscovite) sheared rhyolite grit. This appears to be a sheared epiclastic rock with its dominant constituents, quartz and rhyolite, derived from penecontemporaneous rhyolite such as represented the overlying body.

The extreme northeastern end of the rhyolite body is a medium grey grit containing rhyolite and quartz. Constituents have a weak preferred orientation.

The shape of the rhyolite body south of Fenton Lake, with its gently undulatory western side but strongly convex eastern side suggests that it formed in a depression. It may have been a dome(s) or thick flow(s) which shed clastic debris to the northeast. Part of the body could have been destroyed during eruption or by avalanching after eruption so that the present gently undulatory western side may not be indicative of the original shape of the body.

Southern dome near Dome Lake

This is a prominent dome about 300 m thick at its northern end which tapers out 4.5 km south of column 2 (Fig. 8). Two layers of rhyolite extend northward for about 500 m from the basal and middle parts of the dome. Rhyolite units that occur along the eastern side of Upper Ross Lake and on a peninsula and on islands in the lake, are probably stratigraphic equivalents of the dome near column 2.

Contact relations. The gently undulating base (eastern side) of this rhyolite body is conformable with underlying mafic pillow lavas. This conformable relationship persists southward where the rhyolite thins out and marks a contact between the volcanic belt and the overlying sedimentary succession. Near the southern end of the main body, however, mafic dykes and sills obscure the precise relationship with the overlying Yellowknife Supergroup sediments. Precise structural relations between the steep northern end of the dome and the adjacent andesite pillow lavas were not observed. The contact between the upper layer of rhyolite, that extends northward from this dome, and the andesites to the west is a northerly-trending fault, defined by intense shearing of the rhyolite and by a prominent topographic lineation at least 1500 m long that follows the top of this layer and extends northward to the western side of the next rhyolite dome.

Primary structures. In general, unequivocal primary structural features are lacking within this rhyolite body. The unit has a distinct vertical zonation, however, about 700 m south of column 2 (Fig. 8), where it is about 60 m thick. The lower (eastern) 30 m is massive rhyolite, with weakly developed schistosity, that passes westward into rhyolite breccia and then into a lenticular schistose layer about 10 m thick. The breccia contains angular rhyolite lenses from 200 down to 2 cm. The lenticular layer comprises felsic lenses, 1 to 5 cm long, in a dark grey schistose matrix. Although all parts of the unit are cataclastically deformed, the breccia and lenticular layers appear to have been clastic before tectonic deformation. This succession is interpreted as a rhyolite lava flow that grades upward from a massive basal section to a brecciated top.

Tectonic features. Most of this rhyolite body is cataclastically deformed. Intense shearing along the eastern contact produced a tightly packed mass of lenses and slivers, ranging from 1 to 10 mm wide, that have strong preferred orientation. This orientation defines a steeply dipping ($75-90^\circ$) foliation that is about parallel to the lower boundary of the rhyolite and to the trend of flattened pillow lavas below. In most places a platy cleavage, phyllitic sheen or weak schistosity is developed parallel to the tectonic foliation. Parts of the dome contain interlayers of dark grey felsite, up to 4 m wide, in which the darker colour is due to microscopic biotite in the finely comminuted and recrystallized matrix. The apparent layering is a tectonic feature rather than a primary compositional feature.

Lithology. Rhyolite ranges from massive white or buff weathering outcrops to pale yellowish- or greenish-grey

sheared or lenticular rocks that locally display crude colour banding. Rusty weathering gossan zones, locally containing massive pyrite and generally having prominent iron oxide staining, are common along the contacts of the rhyolite with overlying andesites and in deeply weathered brecciated rhyolite that is adjacent to large metagabbroic sills (in the southern parts of the body).

The fresh rock is a medium to dark grey, aphanitic rhyolite containing sparse (1 to 2 %) microphenocrysts (0.2-0.8 mm across) of quartz, albite and rarely potassic feldspar, in a felsitic matrix. The matrix is an irregular granoblastic mosaic containing 80 to 90 % quartz and feldspar (grains 20-40 microns across) in variable amounts of sericite, biotite, and chlorite, that commonly have preferred orientation. Euhedral poikiloblastic tourmaline is present in one specimen.

The most intensely deformed outcrops weather a resinous pale yellowish- or greenish-grey. The rock may have a streaked texture defined by patches of varying shades of grey or yellowish grey. Darker grey colour banding (resembling bedding) in some areas is caused by microscopic biotite and chlorite concentrated in the recrystallized finely crushed matrix of a lenticular rock. The rock comprises closely packed lenses and tapered fine splinters of rhyolite commonly ranging from 2 to 100 cm long (but ranging down to microscopic lenticles) that have a strong preferred orientation. In some outcrops the lenticular cataclastic foliation superficially resembles eutaxitic foliation typical of densely welded tuffs in undeformed rocks. Similarly, the cataclastic texture could be misinterpreted as a volcanoclastic texture in some outcrops.

In thin section the lenticles range in size down to minute streaks 20 to 50 microns wide. Smaller than this, they cannot be distinguished from crystals of the granoblastic crystallization pattern. Commonly, fine lenticles fit together like splinters from a larger fragment. Masses of fine splinters merge into larger intensely fractured fragments. All sizes of fragments have the same granoblastic crystallization pattern. Phenocrysts and some lithic fragments in the rhyolite have microscopic "crystallization trails" that taper out in the direction of the shearing. Fine mica, which is concentrated in the matrix, has strong preferred orientation parallel to the trend of the lenticles.

The upper layer of rhyolite that extends from the north side of the dome near column 2 (Fig. 8) is a breccia that contains large felsic fragments (ranging from 2 by 4 cm to 10-20 cm wide and 50 cm long) in a darker fine clastic matrix that contains much more mafic constituents than the fragments. This mafic matrix contrasts with the other sheared rhyolite in which the finely sheared matrix is felsic, or only slightly darker than, the larger fragments. This rock was probably a rhyolite breccia before tectonic shearing. The layer is interpreted as an apron of breccia derived from the rhyolite dome.

Interpretation. This body is interpreted as a dome or lava complex that evolved beneath the sea, probably in two or more stages, while andesitic pillow lavas and clastic material continued to accumulate along its side. The layers of breccia that extend from the northern side of the dome are probably

crumble breccias, talus aprons or submarine landslide debris that shed off the dome during various stages of its development. There is no unequivocal evidence that any of this rhyolite had a pyroclastic origin.

Northern dome and breccias near Dome Lake

Very little is known about the northern body of rhyolite because only its margins are exposed along the sides of a northerly-trending lake that conceals most of the body. The body is interpreted to be about 1400 m long and 100 m wide beneath the lake between columns 3 and 4 (Fig. 8) and thins to less than 35 m at each end. Andesitic pillow lavas and pillow breccias overlie the rhyolite on its western side and basaltic pillow lavas and pillow breccias underlie the rhyolite on its eastern side. The following features indicate that the body is at least partially fault bound: predominance of intensely sheared and altered rhyolite and abundance of brecciated rocks along prominent topographic lineaments that follow contacts between rhyolite and andesite along the southern 500 m of the body and between rhyolite and basalts at the northeastern side of the body.

Locally pale green to white weathering rocks, that appear to be massive in outcrops, are intensely sheared and recrystallized, microporphyritic rhyolite in thin section. Relict phenocrysts of quartz (now coarsely polygonized aggregates of quartz 0.2-0.5 mm across) and plagioclase (saussuritized) typically taper out in the plane of microfoliation into fine crystalline "crushed trails" to form distinct lenticular structures. The recrystallized granoblastic matrix of quartz and feldspar contains abundant sericite (up to 40 %) and locally, biotite and poikilo-blastic hornblende that has a strong preferred orientation.

The rhyolite body passes northward into an extensive unit of rhyolite breccia. The precise nature of the transition was not observed. Near the northern side of a fault, about 200 m north of column 4 (Fig. 8), the rhyolite breccia thickens to a maximum of 110 m and continues northward for about 1100 m. Between columns 4 and 5 the rhyolite breccia overlies a 120 m thick lens of breccia containing dominantly metabasaltic fragments but also rhyolite fragments that increase in amount upwards. The Webb Lake Andesite that overlies the rhyolite breccia comprises pillow lavas and breccias and locally coarse andesite and rhyolite fragment breccias. At its northern end the rhyolite breccia appears to grade upwards into these breccias.

The rhyolite breccia at the northern end contains about 80 % tightly packed, angular and subrounded fragments of rhyolite and 20 % fine clastic matrix. Where the amount of matrix increases, this breccia grades into arenaceous breccia. The rock is unsorted and lacks internal stratification. The pale grey to buff weathering rhyolite fragments contain sparse microphenocrysts of quartz similar to the rhyolite body to the south. The matrix is dark grey to reddish-grey weathering, fine clastic material containing fine rhyolitic clasts and dark brown microscopic biotite and chlorite. The mafic constituents have preferred orientation parallel to elongation of fragments. Elongate fragments form a steeply plunging lineation.

About 400 m from the northern end of the unit the rhyolite breccia comprises at least 60 m of rhyolite pebble and boulder conglomerate (largest fragments are near the top) which is overlain by 30 m of intensely sheared "massive" rhyolite that forms the top of the succession. The volcanoclastic unit, near the base of the exposure, contains 15 % white rhyolite fragments that have preferred orientation in a medium grey weathering matrix. Fragments are quartz-phyric rhyolite. The clastic matrix contains coarse-sand size fragments of rhyolite in an aphanitic groundmass containing biotite. The fragments are round to elliptical rods that range from 0.5-3 cm in cross-section and have a length to breadth ratio of up to 4:1. Near the top of the clastic unit, however, the rhyolite fragments increase to cobble size and have round to oval shapes. About 500 m south of this locality rhyolite fragments range up to 60 cm across. The steeply plunging lineations, defined by the elongate clasts, lie on a plane defined by preferred orientation of biotite in a matrix that weathers out as fine elongate pits resembling bedding.

The lens of rhyolite-basalt breccia underlying the rhyolite breccia may occupy a fault-bound depression in the underlying mafic pillow lava succession. The abundance of rounded forms and of elongate clasts in this part of the rhyolite unit (in contrast to the dominance of angular fragments farther north) and the higher proportion of matrix suggest that this may have been a rhyolite pebble and boulder conglomerate. The similarity of lithology to the rhyolite domes to the south suggests that this breccia and conglomerate were derived from those domes or from similar domes that are not exposed in the present erosional section. The breccias may be talus aprons or submarine landslide debris that sloughed off the domes and accumulated partly in a small fault-bound depression. The breccias, containing both rhyolite and basaltic fragments, may represent material derived from a fault scarp in the basaltic pillow lavas as well as from collapsing rhyolite domes.

Dome Lake Basalt

The Dome Lake Basalt is the name given to a bell-shaped body of amphibolite that protrudes from the southwestern side of the volcanic belt 3 km northeast of Dome Lake, (column 2, Fig. 8). A thin tongue of metabasalt, which extends northward for about 700 m from the base of this body, interfingers with the upper part of the Webb Lake Andesite.

The body is a pile of crudely stratified volcanics about 300 m thick. Crude layering defined by alternating pillowed and clastic units (up to 30 m thick) is approximately parallel to, or makes a small angle with, the western contact of the formation. The layering is essentially parallel to the plane of flattening of pillows.

The lower half of the northern and central parts of this pile, between columns 2 and 3 (Fig. 8), comprise alternating units of pillow lavas and coarse volcanoclastics, whereas the upper half contains crudely layered to massive lavas capped by pillowed flows. Margins of pillows in the upper pillowed units are highly amygdaloidal, containing calcite and locally quartz fillings. In one place a pillow "breccia" comprises sack-like bodies, widely separated, within a fine hyaloclastite matrix. These rocks are essentially the same as the pillow breccias described by Henderson (1953) and the isolated pil-

low breccias described by Carlisle (1963). The volcanoclastic units comprise rounded to amoeboid subangular fragments ranging from minute sizes up to 15 cm across. The units are nonsorted, generally unlayered (but locally show a hint of layering in the fine clastic rocks) and contain a high proportion of fine clastic material. The fragments are "peppered" with white amygdules that are stretched and elongated in the direction of shearing and flattening of the pillows. Although deformation and metamorphism has obliterated the fine textures of these rocks, the shape of fragments, their distribution, sorting, layering characteristics, and the gradation into isolated pillow breccias, indicate that these are meta-hyaloclastites.

Cliffs at the westernmost exposures display spectacular stretched (rather than flattened) pillows that weather into columnar-like features. Pillows on horizontal surfaces appear undeformed and have equant forms that range from 30 to 70 cm across. On vertical exposures, however, they are elongate spindles (200-300 cm long) that taper at both ends. Ratios of length to breadth of pillows range from 4 to 10:1. Massive lavas, having widely spaced fractures, occur within the pillowed unit.

Pillow lavas and hyaloclastites containing abundant amygdules, in contrast to generally nonvesicular volcanic rocks of the Webb Lake and Cameron River formations, suggest that the Dome Lake Basalt extruded in a relatively shallow marine environment, possibly at a depth of water less than about 500 m (Moore, 1965; McBirney, 1963). The present erosion surface exposes the deformed flank of a volcanic pile that was not only a westward extension of the volcanic belt but also a topographically high edifice whose peak was in shallow water. This edifice may mark the last eruptive centre in this part of the volcanic belt.

Raquette Lake Formation

The Raquette Lake Formation (Henderson, 1985) is a discontinuous unit comprising heterogeneous assemblages of sandstone, conglomerate and carbonate that extend for about 5 km along the western side of the volcanic belt or the Sleepy Dragon Complex between Upper Ross Lake and Raquette Lake.

The following general description is summarized from the more complete description by Henderson (1985). The formation, which has maximum thickness of about 60 m, overlies conformably, and locally interfingers with, mafic volcanics of the Cameron River Basalt. Commonly, basal conglomerates of the Raquette Lake Formation grade into mafic breccias of the Cameron River Basalt. North of Raquette Lake the Raquette Lake Formation is unconformable with the Sleepy Dragon Complex, although in most places the contact is faulted. In the northernmost exposures, the Raquette Lake succession is overlain conformably by thin felsic volcanic units of the Beaulieu Group. The formation is varied along strike but the most common unit is quartzite. Although all units are rarely present in any one area, the general succession is basal conglomerate overlain by quartzite or felsic volcanic detritus, and locally overlain by calcareous mudstone. Generally, conglomerates contain abundant mafic volcanic clasts, minor felsic volcanic clasts, and rare

granitoid pebbles in a carbonate matrix. In one place, where the formation makes contact with granitoid basement, mafic cobble conglomerate is overlain by granitic fragment conglomerate in which the clasts are similar to the deformed granitoids of the basement complex.

Column 1 in Figure 8b shows the Raquette Lake Formation in one place east of Upper Ross Lake. At this location the Cameron River Basalt is a thin unit of sheared amphibolite. The basal conglomerate is a poorly sorted unit containing subrounded to subangular, elongate pebbles, 2-40 mm across, of pale grey felsite in very fine grained amphibolite, that have preferred orientation in a speckled, dark grey to black matrix. Felsites are recrystallized material whose microfelsic relict texture and mineralogy suggest that they are rhyolite to possibly felsic andesite in composition.

The original texture of the matrix has been obliterated by coarse grained metamorphic minerals. The matrix is dominantly poikiloblastic diopside (1-4 mm across) and microcline (up to 1 m) that enclose quartz, microfelsic material, microcline, epidote and sphene. Irregular areas between microcline and diopside contain microfelsic material that has an irregular granoblastic texture.

The conglomerate grades westward through 20 m of calcite-impregnated conglomerate to massive, very fine grained carbonate about 10 m thick. The carbonate is flanked on the west by 10 m of dark grey rhyolite.

The 40 m thick clastic unit comprising pale grey lenticles in a dark grey matrix lies east of the conglomerate. It appears to be a sheared variety of the conglomerate.

At this locality the conglomerate units appear to be derived from a volcanic terrane since no unequivocal granitic material is present.

The significance of the Raquette Lake Formation is that the granitic cobble conglomerates and quartzite evince erosion from the granitoid basement before and during deposition of the volcanic belt. Henderson (1985) visualized erosion from a rising fault scarp at the margin of the sedimentary basin.

Tumpline Lake subarea

The volcanic belt in the Tumpline Lake subarea is different from those in the other subareas in that volcanism was strongly bimodal and the belt comprises almost equal amounts (by area) of rhyolite and basalt (Fig. 7). The belt is divided into three main formations plus a unit of carbonate: (1) Tumpline Basalt, (2) Turnback Rhyolite and (3) Sharrie Rhyolite. Deformation of the volcanic belt is more intense in this subarea than in the other subareas. Generalized stratigraphic relations of formations in this subarea are shown in Figure 12) and in the legend of Figures 2 and 3.

Tumpline Basalt

Definition, distribution and thickness

The Tumpline Basalt is the name given to the succession of mafic volcanic rocks that lies north of Tumpline Lake, forms easterly trending "wings" of the volcanic belt southeast of

Turnback Lake, and that lies between the Turnback Rhyolite and Sharrie Rhyolite near Sharrie Lake (Fig. 13). The formation is dominantly pillowed flows and layered volcanoclastic rocks of basaltic composition but contains an andesite member near Sharrie Lake.

In almost every area the formation is intensely deformed and thus calculations of thickness depend on structural interpretation (see Fig. 2 cross-sections). The formation may be about 1000 m thick near Sharrie Lake and thins northeastward toward Detour Lake. South of Tumpline Lake thickness ranges from about 500 m on the northern and western sides of the volcanic belt, where it wraps around a large pluton, to less than 100 m on the southern and eastern sides where it interfingers with the Turnback Rhyolite. South and east of Turnback Lake, where the formation is intricately folded, thicknesses may be in the order of 400 to 500 m.

Contact relations

The Tumpline Basalt appears to lie conformably on the Turnback Rhyolite east of Sharrie Lake, southwest of Detour Lake, and south of Tumpline Lake, and on the Sharrie Rhyolite west of Sharrie Lake. South of Tumpline Lake the formation is interpreted as a major wedge within the Turnback Rhyolite but locally the two formations interfinger. In this area the apparent multitude of rhyolite units in the Tumpline Basalt is in part due to complex folding. Southeast of Turnback Lake, the Tumpline formation is conformably overlain by the Turnback Rhyolite.

The basalts are overlain conformably by metasediments of the Burwash Formation southeast of Tumpline Lake (with minor exceptions where thin lenses of rhyolite lie along the contact), locally southeast of Turnback Lake, and north and south of Sharrie Lake.

Southeast of Turnback Lake, the contacts between basalt and Defeat granodiorite are intrusive. Although the volcanic belt appears to conform to the pluton, in detail the contact is crosscutting and dykes of granite intrude the volcanics. Southwest of Tumpline Lake, the Prosperous Granite cuts across layering and trend of flattening of the mafic pillow lavas to form a very irregular contact. In this area, granite encloses blocks and screens of Tumpline Basalt and forms numerous dykes in the volcanic rocks.

Deformation of primary structures

Lithic uniformity and lack of distinct marker units in most places makes it virtually impossible to unravel the structure within vast areas of highly deformed pillow lavas. The contact between the Turnback Rhyolite and the Tumpline Basalt east of Sharrie Lake reveals the structural complexity within an anticlinorium. Deformation has flattened or elongated pillows and locally folded and flattened bodies into undulations 1-3 m in amplitude. In places where pillows appear undeformed and equant on horizontal surface, they are drawn out into steeply plunging rods on vertical planes.

Outcrops of the andesite member near the east side of Sharrie Lake give a spectacular display of intense deformation (Fig. 14) in pillow lavas. Here, pillows are drawn out

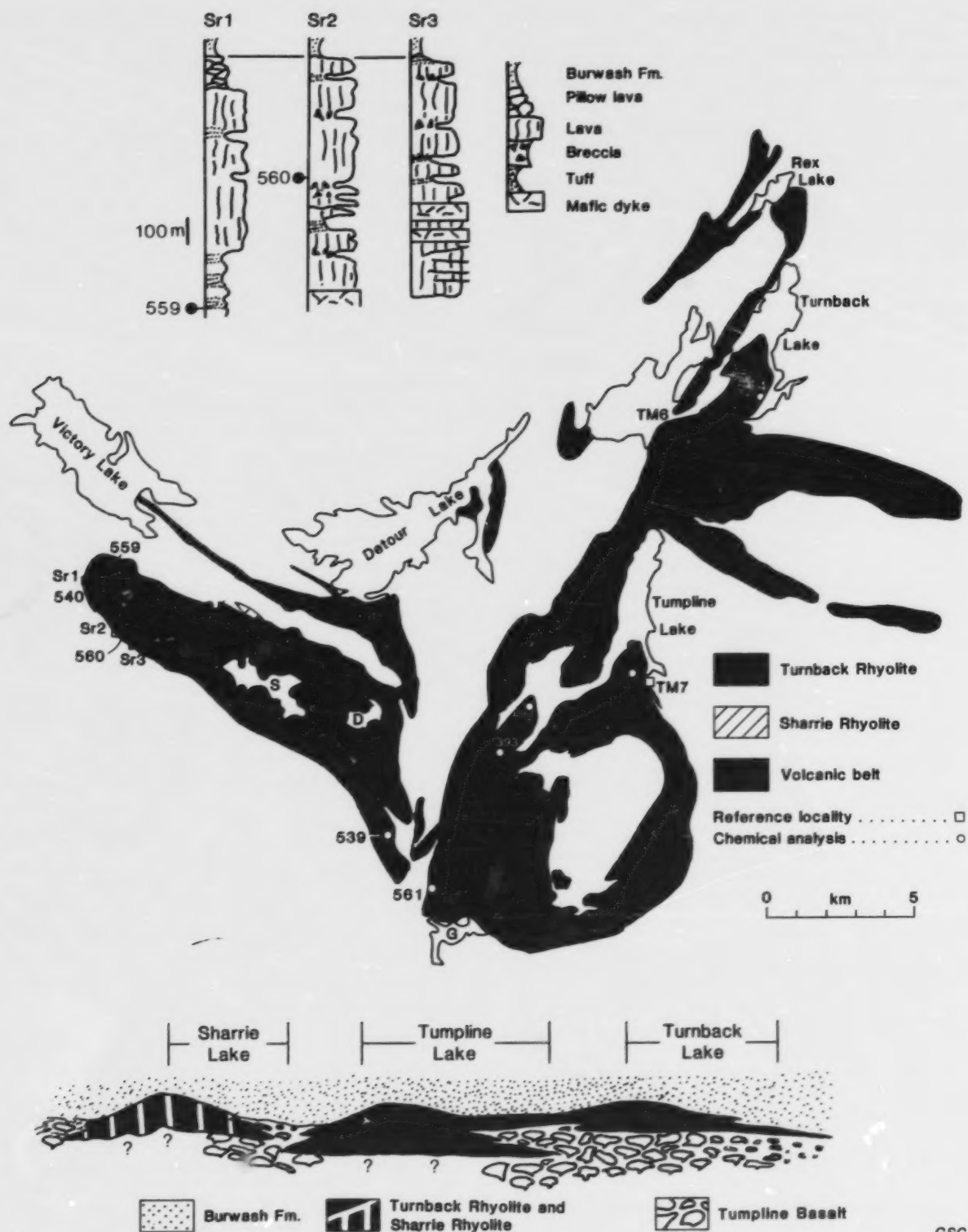
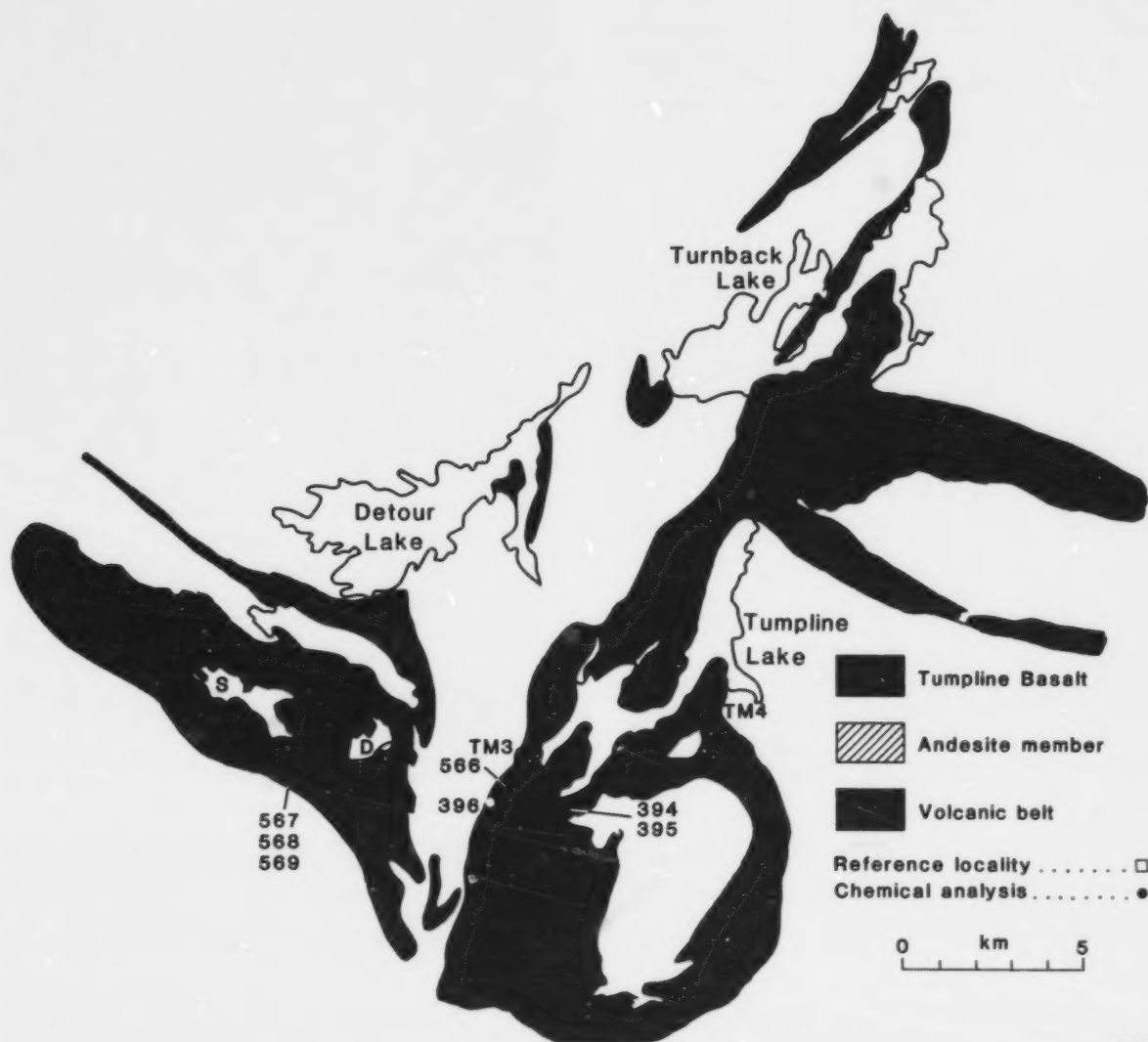


Figure 12. Distribution of Turnback Rhyolite and Sharrie Rhyolite in the Tumpline Lake subarea; reference and chemical analysis localities; S - Sharrie Lake, D - Devore Lake, G - Goose Lake. Columnar sections Sr1, Sr2 and Sr3 show stratigraphy in the Sharrie Rhyolite. Lower section shows schematic stratigraphic relations between rhyolite formations and the Tumpline Basalt.



GSC

Figure 13. Distribution of the Tumpline Basalt; chemical analysis and reference localities. S — Sharrie Lake; D — Devore Lake.

into thin streaks and undulating flattened lenses that taper out to thin or stubby points. Commonly they have digitated, flame-like terminations. Pillows range up to 400 cm long but widths are less than 40 cm. Ratios of length to breadth on horizontal outcrop are greater than 5:1 and locally up to 50:1. In some areas, where pillows are reduced to ribbons 2-3 cm thick, they resemble thin-bedded volcanics. Schistosity, where developed, is generally parallel to the trend of flattening and bedding in the layered successions.

Northwest of Sharrie Lake a columnar jointed intermediate volcanic unit occurs within the mafic pillow lava succession. Undeformed columns have equant, hexagonal shapes in cross-section, ranging from 25-35 m across (Fig. 15a). The joints between columns are lined with white weathering

felsite. Within a single outcrop cross-sections of columns vary from perfect hexagonal forms to flattened shapes that have irregular, wavy and serrated boundaries (Fig. 15b).

Stratigraphy

Turnback Lake area. In the easterly trending wings of the volcanic belt southeast of Turnback Lake, the formation comprises pillow lavas and minor massive lavas and volcaniclastics in the western part (where the wings meet) but it is almost entirely well bedded to crudely layered volcaniclastic rocks in the eastern ends. Massive amphibolites that have intervening zones of mafic breccias and arenaceous clastic material are interpreted as massive lavas.

Layered units at the eastern parts of the wings range from crudely layered rocks to well bedded successions. Some units, characterized by strong differential weathering and lenticular streaks (but lacking distinct beds), are deformed mafic volcanic fragment breccias in which the trend of flattening is almost parallel to gross layering in the succession. Although some boundaries of these layers are sharp, most are indistinct.

Layering in these areas is accentuated by differential weathering along schistosity, which tends to be developed more strongly in some units than others.

Well layered successions are laminated to thin-bedded (beds 0.5-5 cm thick). Beds are straight and continuous on outcrop scale except for local undulations and where they



Figure 14. Deformed pillow lavas in andesite member of the Tumpline Basalt east of Sharrie Lake. (a), (b) and (c) show increasing states of deformation from lenticular bodies having digitated terminations to thin ribbons (c) that have length to breadth ratios about 50:1. Note mylonite zone along right side of (a). Pillows in (d) have felsic rims and more mafic interiors. Areas between pillows are black, micaceous amphibolite containing garnet porphyroblasts. (a) GSC 160335, (b) GSC 160337, (c) GSC 160322, (d) GSC 160333

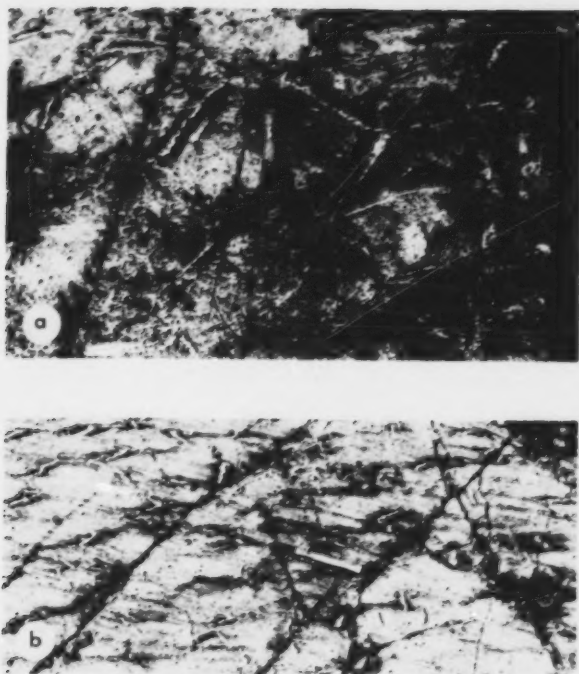


Figure 15. Columnar joints in felsic andesites of the Tumpline Basalt northwest of Sharrie Lake. (a) Undeformed columnar joints. GSC 170828 (b) Deformed joints with crenulated margins, hammer is 33 cm long. GSC 170827

are folded into gentle warps. Boundaries are sharp amongst thinner, finer grained beds and diffuse in thicker coarse grained units. Thin-bedded units are interbedded with arenaceous breccias containing large mafic clasts and lenses. Some units show graded bedding. Generally, clasts are not recognized in the finer laminae because the size of metamorphic crystals in the amphibolite is coarser than the presumed original clasts.

Felsic volcanics of the upper part of the rhyolite formation locally are interbedded with mafic volcanoclastics near the eastern ends of the wings southeast of Turnback Lake. These successions contain alternating thin beds of mafic and felsic volcanics. In one place mafic and felsic materials grade into one another to form compositionally zoned layers.

Tumpline Lake area. Around the large pluton south of Tumpline Lake the formation is dominantly pillow lavas and pillow breccias in the northern parts, mixed pillow lavas and layered rocks along the southern part, and dominantly layered rocks along the southwestern part. All the rocks are intensely deformed so that along the eastern and southern sides some of the "layered rocks" may be intensely flattened pillow lava and pillow breccias (where flat fragments are drawn out into thin ribbons to resemble bedding). That not all the layering is tectonic, is suggested locally by distinct facies transitions from pillow lavas through pillow breccias to layered rocks. Near the contact with the large central pluton some mafic rocks are well foliated schists. Much of the layered succes-

sion is mixed mafic and felsic rocks that represent interlayering on the fine scale of rhyolite and basaltic volcanic sediments. Some of the mafic units contain abundant streaks and lenses of felsic material that represent highly deformed rhyolite fragments.

In the core of a major anticline, about 1 km west of the south end of Tumpline Lake, the formation comprises dominantly layered, to crudely layered lenticular mafic rocks in which deformed pillow lavas and pillow breccias are only locally recognizable. The succession here is interpreted as being dominantly pillow breccias and associated hyaloclastites. Mafic volcanics are fine layered (1-3 cm) for about 75 m from the contact with the Turnback Rhyolite. In some areas this apparent layering is recognized as intensely deformed pillow lavas and pillow breccias with the clasts drawn out into thin streaks. Thus much of the apparent layering may be a result of tectonic deformation rather than relics of primary layering.

Near the contact with rhyolite, southwest of Tumpline Lake (locality TM 1, Fig. 13), intensely deformed pillow lavas grade into fine layered mafic volcanics. This appears to be a facies change rather than a tectonic phenomenon. The gradation occurs where there is abundant mixing of mafic and felsic material near the contact with the rhyolite.

Near locality TM 2 (Fig. 13), south of Tumpline Lake, mafic rocks are almost entirely layered. Generally the mafic volcanics are interlayered with felsic units throughout this part of the belt. Layered rocks range from continuous straight units (1-3 cm thick) that have sharply defined boundaries (Fig. 16a) to interbedded thick- and thin-bedded successions (Fig. 16b).

Sharrie Lake area. In the vicinity of Sharrie Lake the formation is almost entirely pillow lavas except near the eastern extremities of this area where it is layered and schistose amphibolite. The rocks are metabasalts except for folded lentils of felsic andesites west of Devore Lake, 1 km north of Sharrie Lake, and about 3 km northwest of Sharrie Lake. Southwest of Detour Lake the formation is highly deformed, crudely layered amphibolites, commonly containing streaks and lenses of felsic volcanics, and rare pillow lavas. Inclusions of felsic volcanics are common throughout the formation north and east of Sharrie Lake and are particularly abundant near the contact with the Turnback Rhyolite. The mafic formation contains bedded volcanoclastics near the southeastern end of the outcrop area southeast of Devore Lake. Felsic volcanic layers are common within this mafic succession.

Carbonate occurs at the contact between layered felsic and mafic volcanics about 2 km south of Devore Lake. The 20 m thick brown weathering carbonate unit contains abundant quartz veining along bedding.

Siliceous sediments form a unit about 25-60 m wide and at least 1000 m long within the mafic volcanic pillowed succession, about 1 km southwest of Devore Lake. The sediments are a well bedded (1-60 cm thick) succession of poorly sorted plagioclase-quartz wackes. The unit may represent a thin infold of felsic wacke near the top of the pillow lava formation or beds of wacke deposited within the mafic pile.

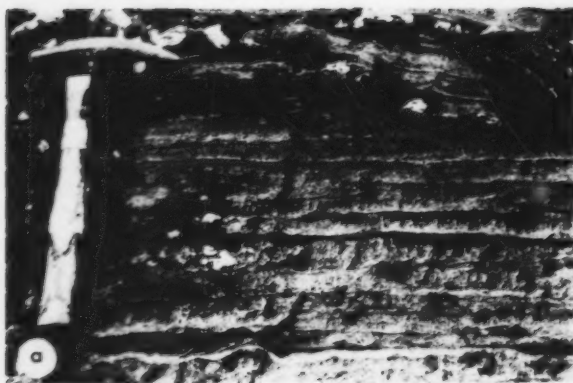


Figure 16. Layered units in mafic volcaniclastic rocks of the Tumpline Basalt southeast of Turnback Lake. These amphibolites are interpreted to have been volcanic wackes and siltstones derived from a mafic volcanic pile to the west. Clastic texture is recognizable only in arenaceous and coarser units. Hammer is 33 cm long. (a) GSC 170749, (b) 170748

Inclusions. Felsic inclusions are abundant throughout the formation in the vicinity of Sharrie Lake, and occur locally near Tumpline and Turnback lakes. The blocks occur both within and between pillows. Fragments are pink, buff, white and very pale green weathering aphanitic felsites that locally have disseminated specks of pyrite. Most appear to be rhyolite or dacite. They range from 1 or 2 cm up to large slabs 10 x 20 m and vary from angular to rounded equant forms, to elongate pods and streaks that are oriented parallel to the trend of pillow flattening. In places felsic blocks that are rounded in cross-section are strongly elongate. Generally they are not as highly deformed as the enclosing mafic pillow lavas. Deformed pillow lavas and pillow breccias mould tightly around the blocks (Fig. 17).

Northeast of Devore Lake felsic inclusions are highly deformed, producing a streaked mafic rock. Similarly, tightly folded layered mafic volcanics west of Tumpline Lake contain abundant felsic material that forms streaks up to 2 cm wide.

Near the contact between andesite pillow lava and Sharrie Rhyolite 1 km north of Sharrie Lake, where the rhyolite thins out between metasediments and metavolcanics, the mafic for-

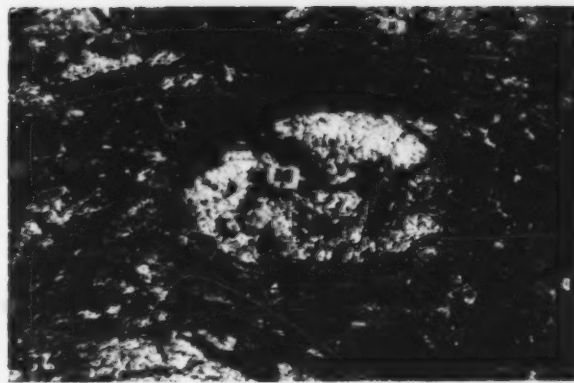


Figure 17. Basalt pillow lavas mould around a felsite block in the Tumpline Basalt northwest of Sharrie Lake. GSC 170824

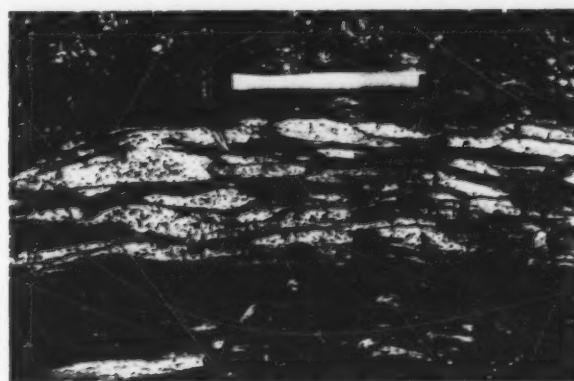


Figure 18. Deformed rhyolite breccia within mafic volcaniclastics of the Tumpline Basalt north of Sharrie Lake. Hammer is 33 cm long. GSC 170829

mation contains zones of fine to crudely layered volcanic arenites that have pods of deformed rhyolite fragment breccia. The breccias comprise intensely flattened rhyolite fragments in a matrix of mafic to intermediate composition (Fig. 18). Blocks in such breccias are derived from the adjacent Sharrie Rhyolite rather than from the Turnback Rhyolite.

Lithology

Basalts. Metabasaltic rocks are dark green to dark grey weathering amphibolites that are dark greenish grey on fresh surface. They vary from fine grained amphibolite to medium grained feldspathic amphibolite to locally schistose biotite-bearing amphibolite. In general, the deformed pillow lavas have aphanitic to fine grained rims less than 1-2 cm thick, whereas the interiors are slightly coarser grained. Very fine grained amphibolites are typically dark green and their outcrops have a smooth glacial polish. Pillows exhibiting vesicular margins, and amygdaloidal cores are rare or not recognizable because of the high degree of deformation and metamorphic recrystallization. At one locality south of Tumpline Lake vesicular pillow lavas are associated with

mixed mafic and felsic volcanic rocks near the contact with the Turnback Rhyolite. Amphibolite with relict phenocrysts of feldspar are rare.

The metabasalts commonly contain 50 to 90 % blue-green hornblende. In some paler aphanitic varieties (dark grey) the amphibole is very fine grained actinolite. Areas between amphibole are invariably a fine, felsitic crystallographic aggregate of plagioclase and quartz. Some metabasalts contain sparse (2-10 %) microphenocrysts of plagioclase (oligoclase to andesine, An_{40}) that are generally less than 1 mm long, and occur as single crystals or glomeroporphyritic clusters. Biotite and epidote occur in minor amounts.

At locality TM 3 (Fig. 13) pillow lavas weather buff to grey and are medium grey to greenish grey on fresh surface. Although the rocks appear to have very little amphibole in hand specimen they have a high specific gravity (about 2.87) in the same range as basalts. Chemistry indicates that these rocks are basalts that have a low normative colour index. Areas between the light weathering deformed pillows are dark grey and amphibole-rich.

Thin- to medium-bedded rocks in the wing of the belt southeast of Turnback Lake contain metasediments interpreted to have been volcanic wackes and siltstones derived from mafic volcanic rocks. Bedding in these rocks is defined by (1) variations in proportion of biotite to hornblende, (2) metamorphic minerals confined to certain layers (eg. garnet), (3) layers of alternating different colour in outcrop related to alternating mafic-rich and mafic-poor beds, (4) changes of coarseness of amphibolite layers, and (5) differential weathering in outcrop. Clastic texture is recognizable only in arenaceous and coarser units because the rocks are almost completely recrystallized.

These crystalloblastic rocks differ from metamorphosed mafic lavas in that they generally contain less amphibole (25-50 %), more biotite (5-25 %), and a higher proportion of felsitic material in which quartz is a major constituent. Biotite-rich sediments commonly contain poikiloblastic garnet making up to 15 % of the rock. Biotite invariably has strong preferred orientation, parallel to bedding and forms a weak schistosity. Abundance of biotite (and locally garnet) suggests that these sediments are more aluminous and potassic than the mafic lavas and dykes.

Volcanic wackes are poorly sorted, containing angular to rounded clasts (0.4-0.7 mm) of quartz, feldspar and crystalloblastic fragments. Elongate grains tend to be orientated in bedding planes. Although clastic texture is rarely preserved in the very fine grained metavolcanic sediments, they commonly contain a few sand sized clasts.

Andesite member. Felsic andesites form a unit about 200 m thick at the east end of Sharrie Lake. Pillows weather pale to medium grey and the fresh surface is dark grey to greenish grey and aphanitic. Pillows margins are more felsic than the interiors (Fig. 14d). They are relatively thick (2-4 cm) in spite of the high degree of deformation. Areas between pillows and pillow rims are dark green, fine- to medium-grained, micaceous amphibolite containing euhedral porphyroblasts of garnet (2-5 mm across) in a mass of black amphibole and

interstitial biotite. Preferred orientation of biotite defines schistosity between pillows.

Felsic margins comprise 5 % microphenocrysts of euhedral plagioclase (less than 0.5 mm long) in a fine irregular crystalloblastic mosaic of quartz and feldspar (grain size ranging from 0.01 to 0.025 mm) that contains 3 to 5 % flakes of hornblende. Plagioclase phenocrysts have preferred orientation. Polygonized relics of plagioclase microlites are rarely distinguishable in the matrix. Lenses and short streaks of coarsely polygonized quartz parallel the feldspar orientation.

Centres of these pillows also have plagioclase microphenocrysts, but contain about 25 % blue-green poikiloblastic hornblende and the microfelsic matrix is a coarser (up to 0.05 mm) crystalloblastic mosaic (than in pillow margins) in which relict microlites are easily identifiable.

Margins of some pillows comprise a series of distinct layers (1 to 2 cm thick) weathering pale grey, greenish grey to brown. One pillow, for example, has a brown outer rim 10 mm thick (comprising 75 % biotite, 8 % hornblende, 17 % plagioclase and quartz) that passes inward through a 3 mm thick hornblende-rich zone into a 10 mm thick felsic zone (containing 2-10 % hornblende, 10 % calcite, 80-90 % quartz and feldspar) then into a series of diffuse hornblende-rich zones 10 to 20 mm thick containing variable amounts of biotite.

A lentil of felsic pillow lavas, thought to be dacite in the field, occurs at the nose of a fold 4 km northwest of Sharrie Lake. The pillows are well formed to stretched bodies up to 1.5 by 3 m. They weather pale grey to pale green and have dark green biotite and hornblende-rich selvages. These rocks may be felsic andesite.

Altered andesites (felsic pillows). At locality TM 4 (Fig. 13) south of Tumpline Lake a lentil of felsic pillow lavas, pillow breccias and hyaloclastite appears within the metabasaltic pillow lava succession. Contacts with the surrounding dark grey mafic volcanics vary from sharp to gradational over a distance of 1.5 m. Pillows and blocks weather white, buff, and pale grey. Dark brownish-green rims are more mafic than the interiors of the pillows. The mafic areas between closely packed pillows are 2-6 cm wide, suggesting that the mafic rims were about 3 cm thick. Pillow rims and matrix between blocks in pillow breccia are garnet-biotite schist resembling a mafic metasediment. Interiors of pillows have a thick white weathering siliceous margin that grades inwards over a distance of at least 2-5 cm into a slightly more mafic (felsic andesite) core. Some blocks and smaller clasts appear to be almost entirely siliceous. The relative composition of core, siliceous margin and mafic rim (or matrix between the pillows) is reflected in specific gravities respectively of 2.73, 2.67, and 3.03. The white siliceous margin is chemically a rhyolite and resembles granoblastic metarhyolite in thin section. The pillow breccias and clasts are almost entirely siliceous material. One block about 45 cm across has an irregular core almost entirely of garnet.

These felsic lavas are believed to be the result of metasomatism that took place within the pillow unit. During alteration of the lavas, if glassy rims altered to chlorite

and clay minerals, then alkalies and alkaline earths have may have been displaced and migrated inward along with silica to produce the rhyolitic margins. The andesitic core may be close to the original composition of the pillow. Metamorphic differentiation may have accentuated original compositional differences within pillows.

Interpretation

The Tumpline Basalt formed entirely in a subaqueous environment as evinced by dominance of pillow lavas that grade laterally into bedded amphibolites. Near Sharrie Lake, basaltic magma erupted through a felsic volcanic pile (Turnback Rhyolite and Sharrie Rhyolite) and incorporated abundant inclusions. The dominantly layered succession southeast of Turnback Lake and south of Tumpline Lake represents distal aprons of volcanic sediments derived from the main volcanic pile to the west and northwest respectively. Interbedding of felsic volcanic material with mafic volcanoclastic sediments may represent simultaneous erosion from both felsic and mafic sources, or deposition of waterlaid rhyolitic material (? tuff) that erupted while the mafic edifice was being degraded. Near Sharrie Lake andesite lavas effused with the renewal of mafic volcanism that followed rhyolitic volcanism. The Tumpline Basalt represents thick piles or possibly ridges of subaqueous pillow lavas that probably erupted from linear vents along a major fracture system. Although it is tempting to relate such a fracture system to the margin of a large uplifted basement block to the north, the high degree of deformation in supracrustals and uncertainties of tectonic interpretation in this area make this suggestion speculative.

Turnback Rhyolite

Definition, distribution and thickness

The Turnback Rhyolite, named after Turnback Lake, includes all rhyolite in the Tumpline Lake subarea that lies directly above the Tumpline Basalt near Turnback Lake and above and below the basalt south of Tumpline Lake where the basalt is interpreted as a wedge within the rhyolite (Fig. 12). The Turnback Rhyolite is stratigraphically the highest formation of the Beaulieu Group in part of this subarea. It makes up about 40% of the volcanic belt and comprises intensely deformed and metamorphosed lavas, (?) domes, and volcanoclastics that probably formed a series of coalescing piles along the margin of the volcano-sedimentary basin.

The formation is best exposed in the noses and crests of large-scale isoclinal folds near Rex Lake, along a line from the eastern peninsula of Turnback Lake down to the southern and western sides of Tumpline Lake, and in areas of rhyolite in contact with the Burwash Formation north of Goose and Cleft lakes.

Thickness of the formation can only be estimated because of thickening and thinning of the unit in hinges and limbs of the tight isoclinal folds. Maximum thickness of the formation is not known. It may be as much as 1000 m south of Tumpline Lake. Near reference locality TM5 (Fig. 12) numerous apparent units of rhyolite within mafic volcanics are synclinal keels of a single unit that is tightly folded. The unit thins eastward from up to 250 m (1 km northeast of the

north end of Tumpline Lake) to 10-25 m in the eastern end of the easterly trending wings of the volcanic belt.

Contact relations

The Turnback Rhyolite lies both above and below the Tumpline Basalt because of large-scale interfingering. Between Turnback and Tumpline lakes, rhyolite appears to lie conformably above the basalt as evinced by (1) parallelism of the contact to layering in both rhyolitic and basaltic formations and (2) local interlayering of mafic and felsic clastic units near the contact. Near the southwestern end of Tumpline Lake, layered rhyolite makes a sharp conformable contact with fine-bedded to crudely layered mafic volcanoclastics. Locally, the two units are interlayered. In many places south of Tumpline Lake the contact is a mixed zone in which layered mafic and felsic volcanoclastic rocks are interbedded. Transition from dominantly felsic to dominantly mafic rocks occurs in one place over a distance of 30 m.

West of Devore Lake, where the rhyolite is overlain by intermediate to mafic pillow lavas, subtle, crude layering in felsic volcanics, and layering and direction of flattening of pillow lavas, all have the same trend, suggesting a conformable relationship. At one locality, a unit of mafic volcanic arenite about 50 m thick lies between the felsic volcanics and pillow lavas. Locally, thin beds and lenses of felsic volcanics interfinger with the overlying mafic rocks.

The base of the rhyolite is not exposed near Tumpline and Devore lakes.

The contact between the rhyolite and the overlying Burwash Formation is conformable in almost every locality, and contacts dip 80-90° because of isoclinal folding throughout the area. Between Turnback Lake and Rex Lake, a 150 m wide carbonate-rich band marks the contact between the rhyolite and Burwash Formation. This band is carbonate impregnated breccia and grit within the margin of the rhyolite. At the south end of Rex Lake the carbonate unit occurs within the rhyolite and not at the sediment-volcanic contact.

At one locality near the north side of Goose Lake, the contact between the Turnback Rhyolite and the Burwash Formation could be interpreted as a structural unconformity by virtue of (a) the high angle between the general trend of foliation and layering in the rhyolite and the volcanoclastic-sedimentary contact and (b) abrupt change of thickness of the rhyolite where it ends against the sediments. Alternatively, the thickness change may reflect a primary abrupt thinning of the rhyolite unit. This critical area was not examined.

The Prosperous Granite makes sharp intrusive contacts with the rhyolite and includes slivers of the rhyolite between Devore and Goose lakes. The Defeat granodiorite south of Tumpline Lake has intruded the rhyolite. Although volcanic units appear to conform to the margins of the pluton on the map, in detail the granodiorite contacts are irregular and cut across layering in the felsic volcanics.

Amphibolite sills generally follow layering in the rhyolite south of Tumpline Lake and north of Goose Lake.

One kilometre northeast of Devore Lake a complex zone, about 200 m wide, of dyke injection occurs between rhyolite and mafic volcanics. There are two episodes of dyking: (1) amphibolite dykes that include clots of felsic volcanics; and (2) rhyolite dykes (up to 10-20 m wide) that swarm through the mafic dykes and felsic volcanics that include large screens of amphibolite.

A swarm of amphibolite dykes intrudes the rhyolite southeast of Detour Lake. A zone of mixed felsic volcanics and fine layers and lenses of mafic material occurs in the felsic volcanics along the margins of the large dykes. These bodies of mixed rock are interpreted as zones of fine injection of mafic material into the rhyolite unit, resulting in layers (0.3-1 m thick) of amphibolite within the rhyolite. Rhyolite and amphibolite near these border zones have intense sulphide alteration. Such gossan zones have been the locus of mineral exploration in the past, as evinced by many blasting pits and trenches.

Stratigraphy

Lavas, breccias, and pyroclastic and epiclastic rhyolites are all recognizable in this formation, but exact facies relations and internal stratigraphy have not been established in most areas. Intense regional deformation and metamorphic recrystallization obscure fine textures and structures that are critical for rigorous description and interpretation. In places, recrystallized textures and structures formed by primary volcanic processes and tectonic processes are difficult to distinguish.

Lavas. Rocks interpreted as lavas occur on the eastern side of the peninsula on Turnback Lake, along the western side of Tumpline Lake, along the southwestern sides of the annular outcrop area around the large pluton south of Tumpline Lake, south of Detour Lake and northwest of Devore Lake.

Lavas are massive to flow-layered bodies, some of which are flanked by breccias. Flow layers, 1 to 10 mm thick, distinguished by different shades of grey, vary from regular laminations, continuous for a few decimetres, to tortuous contortions and small scale folds.

Minor folds of tectonic origin, common in layered rhyolite, differ from the irregular primary contorted layers by forming tight folds of fairly constant orientation that have axial plane cleavage and schistosity. Southwest of Tumpline Lake, for example, S-shaped tight minor folds on a scale of 0.3 to 4 mm are ubiquitous (in both lavas and bedded rocks).

Brecciated margins of rhyolite lavas define large scale layering in some outcrops southwest of Tumpline Lake.

At locality TM6 (Fig. 12), on the east peninsula of Turnback Lake, massive rhyolites that intrude and surround mafic pillow lavas are dykes, sills and possibly domes.

South of Detour Lake, massive, flow layered and coarse block breccias are interpreted as felsic lavas or domes with brecciated tops.

In one area south of Tumpline Lake, massive rhyolite lava grades into coarse breccias that are associated with beds

of coarse and fine clastic material, some of which is conglomeratic. These rocks are flow breccias, talus aprons and epiclastic material derived from the lavas.

Volcaniclastic rocks. The majority of the Turnback Rhyolite contains volcaniclastic rocks including ash-flow tuffs, bedded tuffs, breccias and arenites. Massive volcaniclastics, including nonwelded and partly welded ash-flow tuffs, occur where the felsic volcanic succession is thickest, namely: (1) on the north and western sides of the east peninsula of Turnback Lake, (2) at the west and southwest end of Tumpline Lake, (3) along the northern and western sides of the large pluton south of Tumpline Lake, (4) northwest of Goose Lake, and (5) southeast of Devore Lake. These areas contain rare beds of arenaceous and rudaceous rhyolite between the massive units.

Bedded tuffs and rhyolitic arenites dominate the rhyolite successions in the northeastern, eastern and southern parts of the annular volcanic belt south of Tumpline Lake and form units interbedded with mafic volcanics in the wings of the belt northeast of Tumpline Lake.

On the eastern peninsula of Turnback Lake massive and lenticular volcaniclastic rocks form units up to 600 m thick in which no internal boundaries are recognizable. These thick nonbedded units that grade in and out of eutaxitic zones are interpreted as nonwelded to partly welded ash-flow tuffs. Eutaxitic foliation is defined by preferred orientation of dark grey, aphanitic lenticles, some with fiamme structures on their ends. Recognizable lenticles are commonly 3-5 cm long and about 1 cm wide and, locally, up to 25 cm long and 15 cm wide. The largest lens encountered is 45 cm long and 2 cm wide. In some cases relics of lenticular structure are suggested only by colour mottling (variations in shades of grey) on the weathered surface. Locally, eutaxitic foliation is folded into small-scale (15-25 cm amplitude) S-folds. The tuffs are sparsely porphyritic (phenocrysts of quartz and plagioclase) rhyolite containing lapilli of rhyolite in an aphanitic matrix. Some highly vesicular blocks of pumice, up to 20 cm long and 7 cm wide, are rounded to oval (Fig. 19).

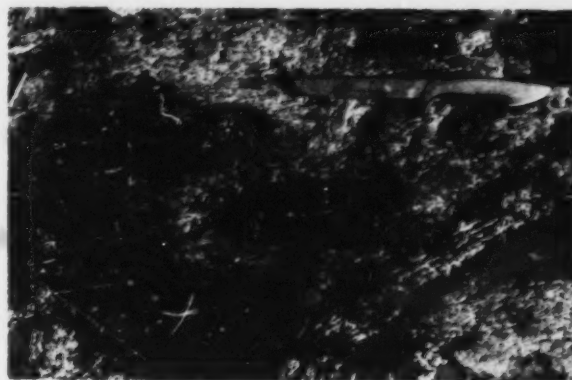


Figure 19. Rounded pumice blocks in ash-flow tuffs of the Turnback Rhyolite on the southeast side of Turnback Lake. Knife is 20 cm long. GSC 170698

Near the southwest end of Tumpline Lake rhyolites contain alternating zones of massive and lenticular rocks, suggestive of zonal welding in ash-flow tuffs (Smith, 1960). Units of fine-layered (1-3 mm) rhyolite that form straight continuous bands across outcrops are sparse within the massive rocks. These well bedded units are regarded as air-fall tuffs within a succession of ash flows.

Rhyolite forming the northwest-trending band of felsic volcanics south of Detour Lake contains a high proportion of layered rocks. Layering represents both volcanic sediments and flow-layered lavas. Near the southern side of this band the felsic volcanics are interbedded with pale grey weathering metasediments. The metasediments are micaceous, fine arenites or lutites that form massive, thick- to medium-bedded successions that tend to be more siliceous than the greywackes of the Burwash Formation. White to very pale grey weathering felsic volcanic units are layered rocks. Apparent layers, ranging from 2 to 14 cm thick, are defined by zones of pitting (less than 1 cm wide), forming slightly curved to linear recessions between massive rhyolite. Some of these layered rocks are broken up into very coarse block breccias. This succession may represent felsic lavas with brecciated tops and intervening volcanoclastic sediments.

Where the band of felsic rocks thins northwestward, silicic rocks commonly have rusty-brown weathering zones (containing abundant finely disseminated pyrite and some pyrrhotite), mainly around mafic units in the felsic volcanics. The felsic volcanics are interbedded with mafic volcanics (dark green sulphide-rich amphibolite) on the northern side and with metasediments (garnet-mica schists) on the southern side.

That the felsic volcanics south of Detour Lake were deposited in a subaqueous environment is suggested by their position between mafic pillow lavas, pillow breccias and meta-arenites to the north, and greywackes and mudstones of the Burwash Formation to the south. These dominantly layered felsic volcanics are probably bedded, waterlaid volcanoclastics, volcarenites or volclutites, and, possibly, waterlaid ash related to the nearby thick successions of felsic volcanics near Devore Lake.

In the wings of the belt northeast of Tumpline Lake bedded rhyolites are interlayered with mafic volcanics, including pillow lavas, massive breccias and crudely bedded volcanic arenites and breccias. Felsic volcanics form units 30 to 100 m thick that are massive to well bedded. Well bedded rocks contain beds ranging from 20 cm to 6 m thick. In places these beds have distinct, fine clastic texture, suggesting that they are volcarenites and lutites that may be waterlaid tuffs or epiclastic rocks.

Intimate interbedding of felsic and mafic units is common in the tightly folded area between 1 to 3 km southwest of Tumpline Lake. At this locality felsic material forms thin layers and lenses throughout the mafic succession that lies between the major areas of rhyolite. In places the transition from felsic to mafic volcanics is a zone of interbedding, about 30 m wide, showing all gradations between dominantly mafic to dominantly felsic material. Layers are generally less than 1 m thick and range down to laminations

a few millimetres thick. Thin layers and lenses in these mixed zones are intensely flattened, stretched and locally boudinaged.

Most of the southern and eastern parts of the annulation south of Tumpline Lake comprise thin- to medium-bedded rhyolite intimately interlayered with mafic volcanics. Rhyolite units 5 to 30 m wide commonly contain layers 1 to 10 cm thick, some of which show normal graded bedding. The dominance of interbedded felsic and mafic volcanoclastic rocks through this area suggests that the succession represents sediments, possibly of both epiclastic and pyroclastic origin, derived from a composite volcanic pile.

Near locality TM7 (Fig. 12) east of Tumpline Lake, metasediments of the Burwash Formation (along the contact with the volcanic belt) contain blocks of both rhyolite and amphibolite (metabasalt), commonly about 25 cm but ranging up to 50 cm long. Blocks are flattened, stretched and commonly have ends that are rounded and digitated. Adjacent mafic pillow lavas are also intensely deformed. These breccias are interpreted as coarse debris that sloughed off the volcanic pile during sedimentation of wackes of the Burwash Formation.

Lithology

Metarhyolites of the Turnback Rhyolite are remarkably uniform through the formation. This apparent uniformity may be due in part to the effects of deformation and metamorphic recrystallization, which tend to alter or obliterate delicate primary textures by which massive rhyolites are distinguished from fine grained pyroclastic rocks. Although the formation has not been subdivided into discrete units, 5 main rock types are recognized: massive porphyritic rhyolite, breccias, tuffs, cataclastic rhyolite, and altered rhyolite.

Rocks of the Turnback Rhyolite weather buff to pale grey, except near contacts with granitic intrusions (Prosperous Granite between Goose and Devore lakes) where they weather salmon pink. Fresh rocks are medium- to dark-grey rhyolites and dacites containing microphenocrysts of quartz and albite in an aphanitic matrix.

Welded ash-flow tuffs (the example described here is from northwest of Devore Lake) feature lenses that have undulatory boundaries that mould around some crystal fragments. Their ends taper into thin wisps or digitate to form a flame structure. Smaller relict clasts (3-5 mm) are equant to lenticular and have irregular, cuspidate boundaries. Matrix to the lenticles contains very fine diffuse lenticles that have undulatory trains of very fine grained mica weaving between them. In thin section the lenticles comprise abundant phenocrysts of quartz and albite set in a microfelsic, crystallographic aggregate of quartz and feldspar, with grain size in the range of 0.01-0.03 mm. The grain size in the matrix to the lenticles is generally coarser (0.03-0.2 mm) than in the lenticles.

In pyroclastic rocks, phenocrysts and crystal fragments of quartz and feldspar are present in both lenticles and matrix and make up 20-25 % of some tuffs. In general, crystals within lenticles are whole, euhedral to subhedral

phenocrysts, whereas in the matrix they are dominantly rounded to irregular fragments of phenocrysts.

Phenocrysts of quartz make up 1-10% of the average rhyolite (up to 20% in some tuffs). Sizes generally range from 0.2-2 mm, and rarely up to 4 mm. Habit varies from euhedral (rare) and subhedral embayed forms (most common) to rounded anhedral forms. In some deformed rocks, quartz forms elongate ovals that have preferred orientation parallel to metamorphic or cataclastic foliation. In relatively undeformed rocks they are mostly single grains that have strong laminar undulose extinction, whereas in deformed breccias they are partially or completely polygonized aggregates. Partly polygonized grains have irregular cores encircled by granoblastic margins.

Phenocrysts of plagioclase form 1-12% of the rock (commonly less than 5%) and generally are less abundant than quartz by a factor of 6 although in some lavas, plagioclase is twice as abundant as quartz. Plagioclase, occurs as single crystals, ranging from 0.2-2 mm and rarely up to 3 mm, or as glomeroporphyritic clusters in some lavas. Crystal habits vary from euhedral laths and rectangular forms to subhedral, rounded and blocky outlines. Boundaries are sharp to minutely irregular where they are intergrown with feldspar of the matrix. Crystals are polysynthetically twinned according to albite and pericline laws and, occasionally, Manebach law. Two-unit broad albite twins are common. Phenocrysts are invariably albite ($An_{0.7}$), have even extinction, and are not zoned. Some, however, have a motley pattern defined by irregular patchy segregations of microcline. Poikilitic grains (usually including quartz) are rare. Extremely fine turbid alteration clouds most of the crystals, except for a narrow margin which is clear.

Matrix to the phenocrysts is a microcrystalline aggregate composed dominantly of crystallographic quartz and feldspar, commonly with minor amounts of biotite, muscovite, chlorite, calcite, rarely epidote, and trace amounts of opaques (pyrite, magnetite, hematite). Grain size of the aggregate may range from 0.01-0.08 mm, but most commonly is 0.03-0.05 mm. Texture of the felsic aggregate varies from heteroblastic to granoblastic overprinted by lepidoblastic micas. In lavas, dykes and sills, the texture may be mottled, where relict pyroclasts or rhyolitic lithic fragments have a crystallization pattern slightly different, in grain size or uniformity of grain size of the felsic aggregate grow, the surrounding matrix.

The Prosperous Granite between Goose and Devore lakes apparently has altered rhyolites along the intrusive contact. Rhyolites at the contact weather salmon pink and contain abundant veins of quartz that lie along foliation in the volcanic rocks. Quartz veins range from minute stringers to bodies 40 cm wide. In some areas the rhyolite has a lenticular foliation that is parallel to the intrusive contact, which cuts across the trend of the volcanic units at a high angle. The lenticular structure suggests that pyroclastic rocks adjacent to the granitic contact may be relict fused tuffs.

Textures of the rhyolite near Rex Lake vary from aphanitic to very fine grained with a sugary texture and a schi-

stose structure. Generally, the grain size and the amount of muscovite increase in the rhyolite towards the large body of Redout Granite and the halo of small satellite plutons that intrude the rhyolite. In places it is difficult to distinguish the sugary textured rhyolite from the very fine grained, chilled margin of the granite. In most places, however, the rhyolite is well layered. The coarsening of grain size and the increase in mica is regarded as a contact metamorphic effect on the rhyolite.

Recognition of pyroclastic rocks. Recognition of the pyroclastic origin of rhyolites in this formation is based on patterns of textural and structural change within clastic units on outcrop scale and on interpretation of relict textures in thin sections.

Distinct clastic texture is present in some areas, but is commonly not obvious. Clastic textures visible in hand specimen commonly are extremely subtle and difficult to identify in thin section. Matrix of all rock types of the Turnback Rhyolite are microfelsites. Generally only the relics of coarse (lapilli-sized) pyroclastic particles are preserved, whereas ash-sized material is rarely recognizable because of the coarseness of the crystalloblastic groundmass. Large amounts of crystal fragments and phenocrysts in the massive rocks are considered as supporting evidence that the rocks may be pyroclastic in origin.

In general the main lithological variations in massive rocks are the abundance of phenocrysts and the degree to which lenticular textures are developed.

During initial stages of devitrification and recrystallization of primary vitric tuffs, relatively coarse glassy fragments (i.e. pumiceous material), commonly develop coarser (or in some cases finer) crystalline aggregates than the fine shards of the ashy matrix (Lambert, 1974). Metamorphism results in breaking down and coalescence of grains in the initial aggregate to form coarser grained aggregates. The initial difference in crystallization pattern between pumice and shards is reflected in contrasting coarseness of crystallization in the metamorphosed rock. Hence areas of contrasting crystallization may define relics of pumice, shards and cuspidate volcanic fragments, indicating a pyroclastic origin for the clasts.

Features in the Turnback Rhyolite that resemble ash-flow tuffs include: massive character (general lack of layering), subtle to well defined lenticular and streaked features having preferred orientation resembling eutaxitic foliation, general lack of coarse clastic material, and lack of identifiable contacts over broad areas.

Recognition of welded tuffs is a point of controversy in many areas because criteria depend on the interpretation that lenticular structures (or *fiamme*) formed during compaction of hot ash flows. *Fiamme*-like structures can form during post-depositional processes such as hydrothermal alteration, by fusion of a tuff adjacent to a hot intrusion, purely tectonic processes such as shearing and crushing in fault zones, and by tectonic flattening of original equant fragments.

In welded tuffs of the Turnback Rhyolite the degree of deformation varies markedly between different types of fragments. For example, relict pumice lenticles are flattened or deformed around lithic and crystal fragments that are essentially undeformed. Eutaxitic foliation tends to have zonal distribution with respect to the units in which it occurs.

In rocks that have suffered tectonic deformation, all types of fragments show evidence of deformation which tends to be fairly uniform in particular outcrop.

Tectonically deformed rhyolites. Tectonically deformed and recrystallized felsic volcanic rocks may have textures and structures that strongly resemble relics of primary volcanic features. It is critical to distinguish whether fine clastic textures and lenticular structures were formed by tectonic or pyroclastic mechanisms. Incorrect distinction of these features may lead to erroneous interpretation of rocks as being tuffs or ignimbrites.

All rocks have suffered tectonic deformation and metamorphic recrystallization. The rhyolite occurs in cores of huge, tight, isoclinal folds and in units deformed around plutonic bodies. The rocks have a metamorphic foliation defined by phyllitic sheen or a weak to prominent schistosity that is predominantly steeply dipping and essentially parallel to layering, lenticular structure, and to limbs and axial surfaces of isoclinal folds. In some areas metamorphic foliation in rhyolites is recognizable only in thin section by crude preferred orientation of minute mica flakes in the aphanitic matrix. Weathering along this subtle foliation or related cleavage gives massive outcrops a fine pitted surface that may be interpreted, erroneously, to indicate a fine clastic texture. Locally, where trains of fine pits defining a penetrative structure are folded, they resemble subtle primary layering.

Lenticular rocks formed by tectonic deformation and metamorphic recrystallization have the following features: (1) microshears, that undulate and coalesce to produce a mass of lenticular slivers or pinching and swelling areas resembling flattened lenticles, (2) boundaries of slivers (lenticles) tend to be sharply delineated, (3) recrystallized zones, that separate and define lenticles, are delineated by trains of mica, dominantly biotite, and by narrow zones or streaks of granoblastic, quartz and feldspar that is slightly coarser than the crystalloblastic aggregate on both sides of the shears, (4) granoblastic aggregates generally comprise equant quartz and feldspar that have sharply defined boundaries and form abundant triple junctions, (5) granulation and recrystallization around crystals produce granoblastic mortar trails or pressure shadows that taper in the plane of foliation, (6) crystals are pulled apart in the plane of foliation, and (7) crystals in recrystallized shear zones are wholly polygonized whereas those within lenses are subhedral to euhedral, having undulose extinction but generally are not polygonized.

At one locality south of Tumpline Lake deformed breccias contain intensely flattened rhyolite clasts up to 30 cm long and 10 cm wide that have strong preferred orientation (Fig. 20). Some of the larger blocks have flame structures

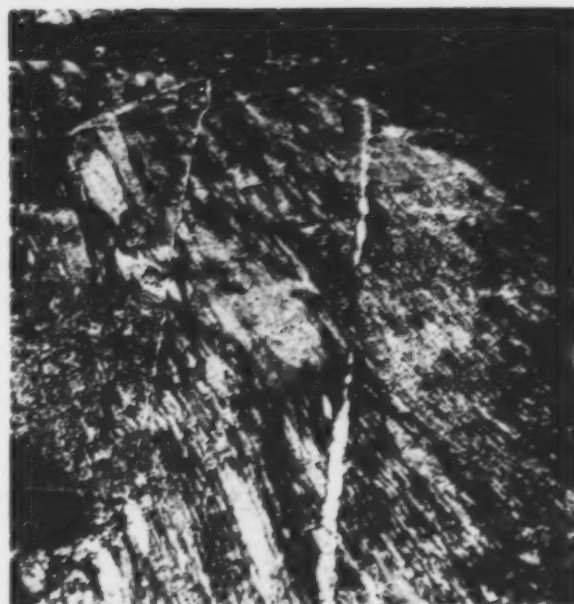


Figure 20. Tectonically deformed rhyolite fragment breccia having pseudo-eutaxitic foliation in Turnback Rhyolite north of Goose Lake. GSC 160298

at their ends, but most taper out smoothly. The strong preferred orientation of the flattened blocks resembles eutaxitic foliation. Similarly, shearing in cataclastic zones may produce a lenticular structure in outcrop that resembles eutaxitic foliation of welded tuffs. Lenticles formed by shearing, however, lack the criteria which commonly distinguish metamorphosed pumice lenticles from their surroundings; i.e. contrasting grain size of crystallographic aggregates; contrasting shape, abundance and population of phenocrysts; contrasting amounts of mafic minerals and irregular, cuspidate boundaries. Primary volcanic lenticles are not bound by microshears or by recrystallized zones of granulation.

In some areas south of Tumpline Lake crystal-rich rocks have crystallization features of tuffs, and eutaxitic foliation resembling welded tuffs, but also features of cataclastic rocks, indicating that both processes contributed to the final texture.

Interpretation

Lava flows are distinguished at a few localities on the basis of flow layering and autobreccias. Some of the coarse breccias near these bodies may be aprons of rubble on the flanks of thick massive flows or possibly domes. No domes, however, have been positively identified.

Some of the massive pyroclastic rocks south of Tumpline Lake and southeast of Devore Lake are ash-flow tuffs. Their massive nature, general lack of intercollated sediments, and presence of welded zones, suggest that at least some were subaerial ash flows. Eutaxitic structures in

tuffs about 3 km north of Goose Lake (that parallel contacts of granitic intrusions), however, probably are fused tuffs rather than ignimbrites. Thus, clasts in the tuff were softened and deformed in response to intrusion of the hot granitic body. Layering and sorting patterns suggestive of subaqueous ash flows, as described by Fiske (1963), Fiske and Matsuda (1964) and others, were not found.

The interlayering of the rhyolite with mafic pillow lavas south of Tumpline Lake and interbedding of layered felsic and mafic volcanic arenites that are closely associated with pillow lavas near Devore Lake and in the wings northeast of Tumpline Lake, indicate that some of the rhyolite formed in a subaqueous environment.

Ash-flow tuffs occur in several areas where the Turnback Rhyolite is thickest. Away from these areas the volcanics are bedded volcanoclastics interpreted as epiclastic sediments and waterlaid tuffs. The areas of thick ash flows are assumed to have been near major centres of eruption, some of which were emergent, whereas the peripheral bedded successions represent epiclastic deposits. The unusually high proportion of felsic volcanic rocks in this subarea could be related to the Defeat Plutonic Suite which forms several intrusions within the volcanic belt. Perhaps the rhyolite volcanism was the first manifestation of the impending granite plutonism which eventually climaxed in a swarm of intrusions.

Sharrie Rhyolite

The Sharrie Rhyolite includes rhyolitic rocks exposed at the nose of the anticlinorium northwest of Sharrie Lake. Northwesterly trending bands of rhyolite exposed along the crests of isoclinal between Detour and Victory lakes are considered to be part of the Sharrie Rhyolite.

The formation comprises sparsely porphyritic lavas, (?) domes, minor pyroclastic rocks and crystal tuff. Maximum thickness is in the order of 850 m near Sharrie Lake. In this area the unit is thickest at the nose of the fold, tapers out on the northern side of the anticlinorium, and fingers out into metabasaltic volcanics on the southern side of the fold.

The Sharrie Rhyolite appears to lie conformably between the Tumpline Basalt and the Burwash Formation. Near the nose and on the southwestern side of the anticlinorium, lenses of mafic volcanics and felsic andesite pillow lavas occur between the rhyolite and the Burwash sediments (column Sr1, Fig. 12). Gossans about 15 m wide are common along both the upper and lower boundaries of the formation.

Figure 12 shows stratigraphy within the formation in three columnar sections (Sr1, Sr2 and Sr3). The succession consists dominantly of lavas and minor pyroclastic rocks, mainly in the lower parts of some sections. Rhyolites near Sharrie Lake show less internal deformation and metamorphic recrystallization than rhyolites in other areas in the Tumpline Lake subarea.

Lava flow units (or possibly domes) range from 40 to as much as 250 m thick. They have massive to flow layered interiors and brecciated tops and bottoms. Clastic zones that

mark the boundaries between flow units vary from volcanic arenites, tuff-breccias (less than 15 m thick) and in one place rhyolite boulder conglomerate, to coarse flow-top breccias that have maximum thicknesses of 20 to 40 m. Breccias marking the boundaries of lava flows comprise a jumble of unsorted angular to subrounded blocks, up to 35 cm across, set in a matrix of sheared rhyolite. Deformed blocks are generally elongate parallel to foliation in the sheared rocks. Breccias that grade into massive or flow layered interiors of flows are regarded as autoclastic flow-top breccias. Some lenticular rhyolite breccias, however, are shear zones within the flows. In general, flows are not highly deformed and most shearing appears to have taken place within breccia zones between flows. A phyllitic cleavage is present in most rocks. Weathering along this cleavage forms trains of elongate pits that give outcrops an appearance of being subtly layered.

White, buff and pale grey weathering lavas are medium grey, sparsely porphyritic rhyolite containing 1-5 % phenocrysts of plagioclase and quartz, less than 1 mm across. Plagioclase occurs as single crystals and as glomeroporphyritic clusters. They have euhedral to subhedral outlines, some of which are embayed. Compositions range from albite to oligoclase (An_{8-15}) but crystals are not zoned. In some rocks they have a crude preferred orientation with the same trends as the flow layering in the rock. Phenocrysts of quartz form subhedral rounded grains. Matrix is a uniform crystalloblastic felsic aggregate containing irregular quartz and feldspar. In one flow, microlites of plagioclase (0.026 by 0.1 mm) are preserved in the matrix. They form subhedral to anhedral laths (boundaries intergrown with the surrounding microfelsic aggregate) that have preferred orientation defining a trachytic texture that flows around phenocrysts.

Pyroclastic rocks occur near the centre of the anticline. In column Sr2 (Fig. 12) a unit of rhyolite tuff about 35 m thick lies between brecciated margins of lava flows above and below. The unit comprises massive arenaceous tuff and thin (about 1 cm) beds of quartz crystal tuff. The massive unit is a medium grey, poorly sorted recrystallized tuff containing 15-20 % rounded to elongate quartz grains (0.5-2 mm) and 2-5 % equant to blocky crystals of feldspar (0.2-0.5 mm) that are randomly distributed in an aphanitic irregular crystallographic matrix. The matrix is a microfelsic aggregate of quartz and feldspar in which equant to lenticular patches of granoblastic crystallization, surrounded by a finer grained irregular crystallographic aggregate, define a relict clastic texture. Since no lithic volcanic fragments are recognized, this unit was probably a vitric crystal ash.

The crystal tuff layer comprises 45 % crystals of quartz and 1-2 % crystals of plagioclase in a recrystallized matrix. Although crystals range from 0.7-2.0 mm, most are 1-1.2 mm and the rock is moderately well sorted with respect to the sizes of phenocrysts. Quartz crystals have subhedral, anhedral and rounded forms, and some are broken. Matrix between the crystals contains coarse (about 0.8 mm) granoblastic plates of microcline and plagioclase poikiloblastically enclosing dense swarms of minute rounded quartz. The inclusions have a preferred orientation throughout the

matrix even through the enclosing grains have random crystallographic orientation. Granoblastic quartz forms irregular areas between the poikiloblasts. Although there are no relics of pyroclastic texture, the high concentration of well sorted crystals and crystal fragments, and uniformity of crystal species, suggest that the rock is an air-fall tuff.

Massive fine clastic units, in the lower 100 m of column Srl (Fig. 12), include rocks having lenticular structure defined by pale grey, elongate lenses up to 5 cm long (one is 15×17 cm) that have strong preferred orientation resembling eutaxitic foliation of welded tuffs. Rocks that are not lenticular are sparsely porphyritic, buff to pale grey weathering rhyolites containing felsic clasts up to 5 cm across in a dark grey aphanitic matrix. In thin section the rocks are completely recrystallized crystallographic microfelsesites in which clasts, that are clearly recognizable in hand specimen, are barely distinguishable from their matrix by slight contrast in coarseness of the crystallization pattern and lesser amounts of mica. These rocks may have been vitric tuffs in which clasts and fine matrix were of similar composition and thus recrystallized to form mineralogically identical aggregates. Only the texture distinguishes clasts from matrix.

In one place the volcanoclastic rock is a poorly sorted volcanic wacke comprising subrounded to subangular grains of quartz and metarhyolite with a wide range in grain size. Elongate quartz grains and lithic fragments have preferred orientation. Porphyroblasts of garnet up to 6 mm across enclose trains of minute quartz grains that are parallel to the mineral orientation of the rock. This rock has a much higher amount of mafic material than most clastic rhyolites (30% hornblende and biotite) and has biotite schistosity parallel to bedding.

Interpretation

Although the Sharrie Rhyolite lies below pillow lavas and turbidites of the Burwash Formation, most units within the formation are not typical of subaqueous deposition but are massive lavas and tuffs. Exceptions are the volcanoclastic units that finger out into pillow lavas on the eastern extremities of the formation and that lie within the Burwash sediments to the north. The formation is interpreted as marking a centre of felsic volcanism that began with explosive activity in a shallow submarine environment. The explosions, probably phreatomagmatic eruptions that broke the surface of the water, distributed pyroclastic material for several kilometres around the surrounding area to be deposited as waterlaid ash amongst mafic pillow lava successions and on sediments accumulating adjacent to the volcanic belt. Quiet effusion of rhyolite magma formed a complex of flows and domes that rose out of the sea to form ephemeral islands.

Sunset Lake subarea

The volcanic belt in the Sunset Lake subarea comprises three formal formations and two informal units: (1) ultramafic rocks (informal) (2) Sunset Lake Basalt, (3) Alice Formation which is divided into two members, an andesite member and a dacite member, (4) rhyolite (informal), and

(5) Payne Lake Formation. Figure 3 shows facies relations in this subarea.

The degree of metamorphism varies within the belt from amphibolite facies in the western and eastern limbs to greenschist facies along the east side of Sunset Lake. The low metamorphic grade in the vicinity of Sunset Lake coincides roughly with the core of a syncline that plunges gently south. Thus the area of greenschist metamorphism may correspond to units that are highest in the volcanic succession.

At least 3 generations of folds have affected this belt (Lambert and van Staal, 1987) so that in most places pillow lavas are greatly flattened and elongated and steeply dipping foliations and bedding are ubiquitous.

Ultramafic rocks and biotite schist

Detailed investigations of boundary relationships between the Sleepy Dragon Complex and the volcanic belt by Lambert and van Staal (1987), led to the recognition of ultramafic rocks in the Beaulieu River volcanic belt which coincided with their discovery in volcanic belts near Cameron River and Point Lake (T.M. Kusky, pers. comm., 1986) and follows the first revelation of ultramafic compositions in the Slave Province, in the Hope Bay greenstone belt, by Gibbins and Hogarth (1986).

Ultramafic rocks were mapped for 3 km along the westernmost edge of the volcanic belt where it is in contact with granitoid gneisses, locally mylonitic, of the Sleepy Dragon Complex (Fig. 3, 21, 22). Micaceous (mainly biotite) schists occur within, but mainly as 20-50 m wide lenses immediately west and south of, the ultramafite. The ultramafite pinches and swells, has a maximum width of 60 m, and tapers out southwards into biotite schist. Its extent in a northerly direction is not known. Contact zones with adjacent dark green schistose mafic volcanics, mafic pillow lavas, amphibolite dykes and granitoid gneisses generally have a strong foliation.

Typically, ultramafic rocks form smooth low relief, pale brown to pale green weathering, whalebacks in areas of poor or intermittent outcrop. The northern part of the unit is poorly exposed and commonly found as thin schistose smears along the western edges of volcanic bluffs that line a small northeasterly trending valley linking a series of lakes. The widest and best exposure of ultramafic rocks on the eastern and southern shore of a small lake at locality A (Fig. 22) shows the complex relationships between amphibolite dyke, granitoid gneiss and ultramafite. The ultramafite is largely a highly deformed serpentinite that shows a wide range of mineralogical and textural varieties including aphanitic serpentinite (mainly antigorite) with veins of chrysotile, fibrous actinolite, talc and chlorite schists, and relict coarse grained rocks. Some rocks contain olivine, enstatite and amphibole (?anthophyllite). Zones of contrasting texture may represent primary igneous layering.

The long narrow outcrop pattern, with orientation similar to amphibolite intrusions in the complex boundary zone, suggest that the ultramafite could be a dyke or sill.

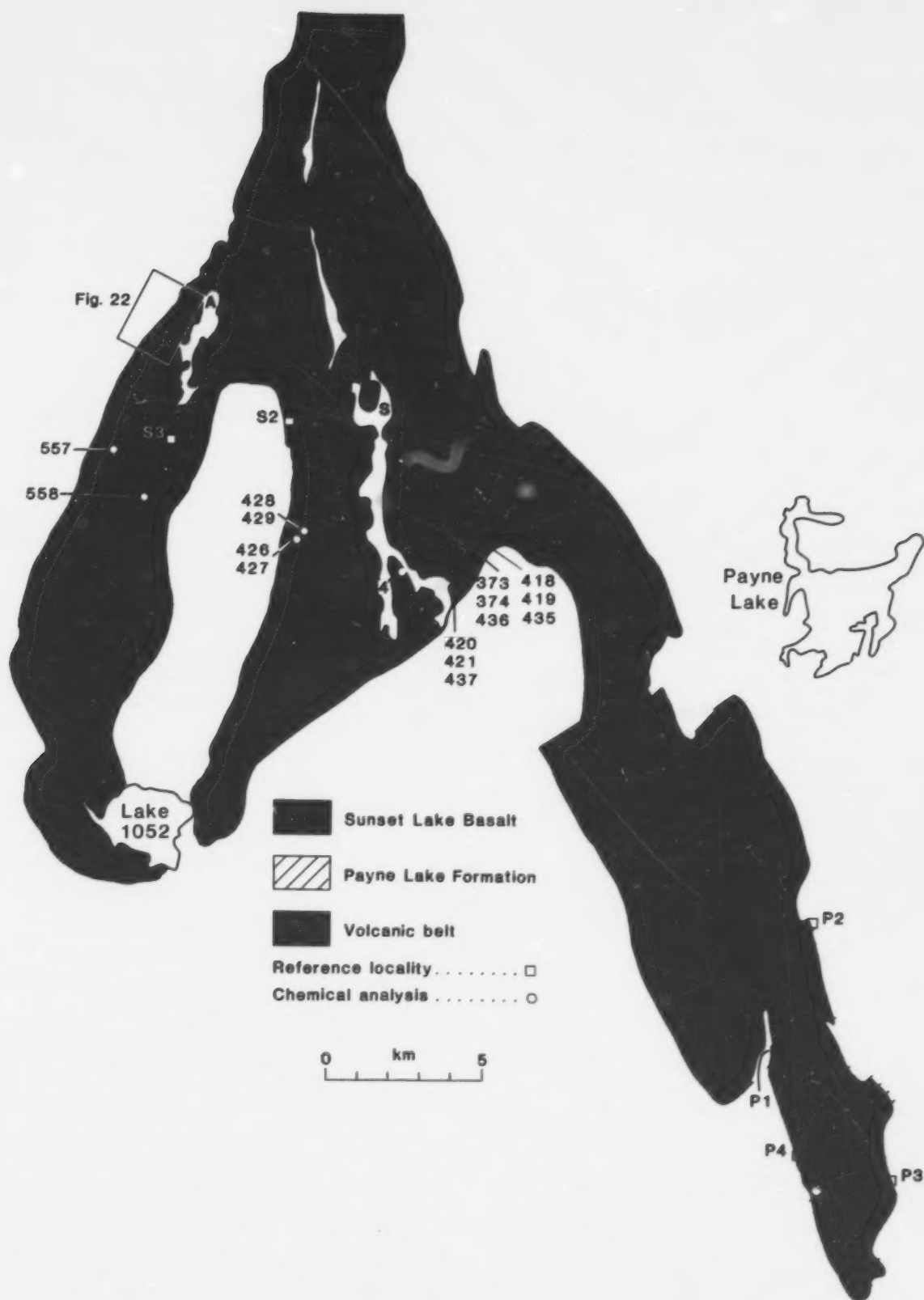


Figure 21. Distribution of Sunset Lake Basalt and Payne Lake Formation; reference and chemical analysis localities. S — Sunset Lake; A — Amacher Lake.

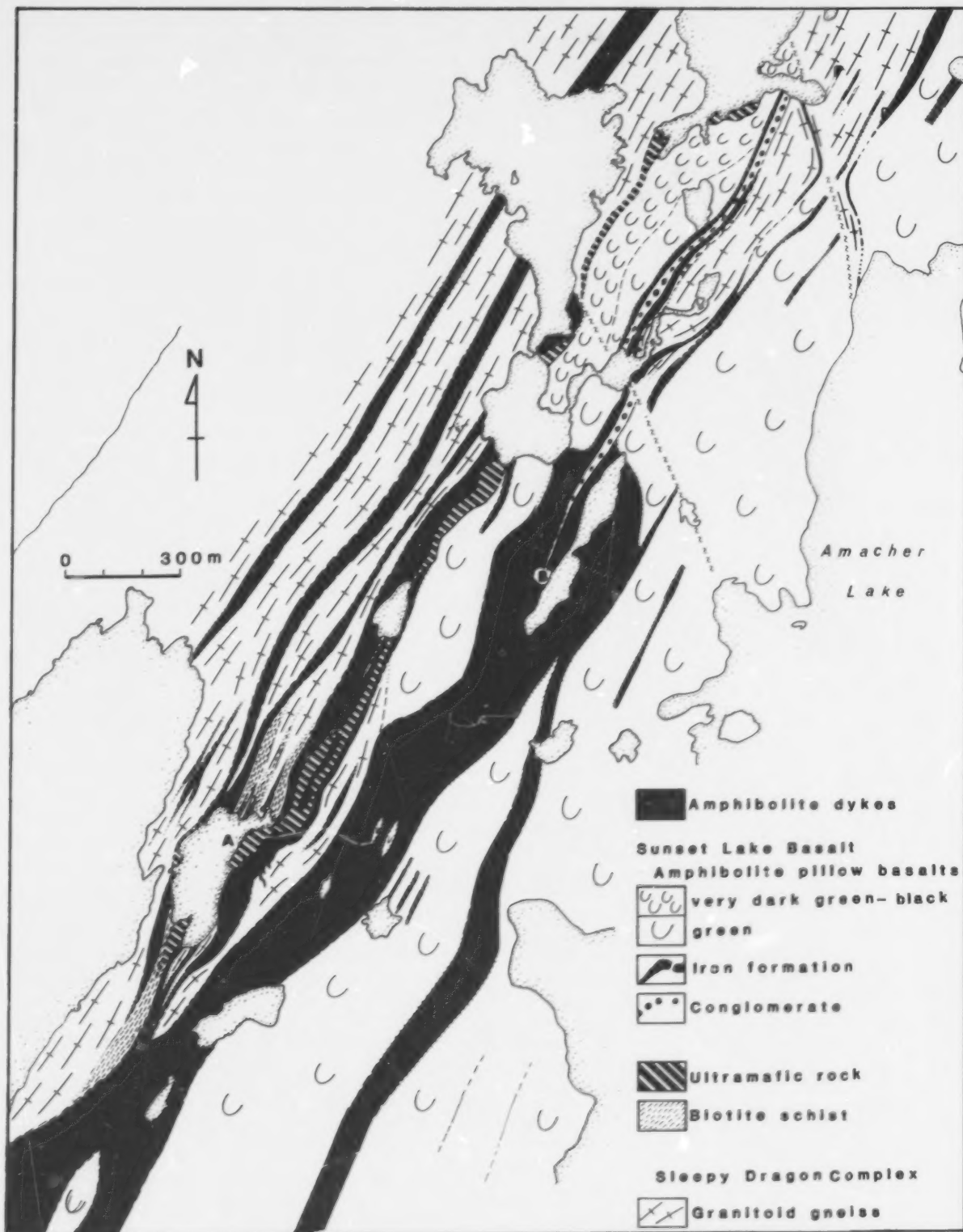


Figure 22. Granite-greenstone boundary zone near Amacher Lake, Beaulieu River volcanic belt. Figure 21 shows the location of this figure (after Lambert and van Staal, 1987).

Sunset Lake Basalt

Definition and distribution

This basaltic formation makes up about 70 % of the volcanic belt in the Sunset Lake subarea (Fig. 21) and forms the oldest and most voluminous deposits in that subarea. It comprises dominantly basaltic pillow lavas, pillow breccias and associated hyaloclastites, but also contains minor iron-formation, conglomerate and volcanic sediments. It lies below or interfingers with the Alice Formation, and rhyolites and volcanoclastic sediments of the Payne Formation. The precise thickness of the formation is unknown.

Contacts

The complex boundary zone between the Sleepy Dragon Complex and the Sunset Lake Basalt along the western side of the Beaulieu volcanic belt, which has been described above, is generally a linear to gently undulating zone of high ductile strain complicated by large amphibolite dykes and the Amacher shear zone (see Structural geology). Granitic rocks, amphibolite dykes and volcanic rocks are all intensely deformed near this boundary, and high shear strain extends up to 1 km into the volcanic belt. Fault bound lenses of granitoid gneiss occur in the Sunset Lake Basalt within 1 km of the main volcanic contact.

Northwest of Payne Lake, where Sunset Lake Basalt forms the northeastern margin of the volcanic belt, it variably has (1) sheared contacts with the mafic dyke swarm-granitoid gneiss complex (see Dykes and sills), (2) intrusive contacts with the Meander Lake Plutonic Suite and (3) sheared contacts with Meander Lake rocks where both the volcanic belt and the plutonic suite have been cut by the Payne shear zone.

Conformable contacts between the Sunset Lake Basalt and the Burwash Formation are exposed between Sunset and Payne lakes. Although shales in the Burwash Formation are deformed at the contact, the general trend of bedding is parallel to the trend of pillow lavas. The contact in this area, indicated by the attitude of bedding in the shales, is overturned steeply to the east. In many areas apparent conformity may be inevitable in the light of the intense flattening and deformation in the rocks.

Northwest of Sunset Lake, fingers of the basalt lie conformably within the Alice Formation. The tongue of basalt pillow lavas, extending about 5 km north of the Amacher Granite west of Sunset Lake, is highly interpretive because of sparse outcrop.

Near the northwestern side of Sunset Lake the basalt is intruded by felsic dykes and sills related to rhyolite domes.

Primary structures

Layering. Layered mafic volcanic rocks are relatively rare in this formation. They form a band about 200 m wide and extending south-southwestward from Amacher Lake for at least 3.5 km; a northerly trending band 200 m wide and at least 2.5 km long about 3 km northwest of Lake 1052; bedded successions south of Sunset Lake near the rhyolite dome

and near the contact with Burwash sediments; and local layered units within the pillow lava succession near the southern end of the formation west of Payne Lake. Layering, in most places bedding in metavolcanoclastic sediments, almost invariably is parallel to the local trends of the volcanic belt and to the trend of flattening of pillow lavas.

West of Payne Lake, mafic volcanics contain layers (about 3 m wide) of layered felsic volcanics. These bodies are interpreted as rhyolitic dykes and sills, but could represent waterlaid tuff related to rhyolite domes at the south end of Sunset Lake.

Pillows and pillow breccias. Except for minor layered successions and rare massive lavas, the entire formation comprises closely packed, deformed pillow lavas, pillow breccias and minor hyaloclastites. Features of pillows are essentially the same as those described in the Cameron River and Tumpline basalts.

At locality S 2 (Fig. 21) mafic (almost black, fine grained amphibolite) pillows display radial fractures, a feature rarely observed in pillows in any of the subareas. Radial fractures are best developed in the outer third of pillows, whereas fractures are more irregularly oriented in the core.

Pillow breccias in many places comprise angular, rounded, equant and elongate clasts and parts of broken pillows in a fine to coarse unsorted clastic matrix. In places, however, pillow breccias are a mass of strongly oriented, rounded ovoids and lenses that have smooth, rounded terminations and thin rims. Angular fragments are rare and most fragments have globular forms. The matrix is fine clastic to massive amphibolite that represents metahyaloclastite. Large pillows are sparse within the breccia.

Massive lavas. Massive flows within the pillow lava succession are fine grained, medium green amphibolite. A zone of intense pock-marking that grades into one unit but makes sharp contact with another commonly marks the boundary between two flows. Such zones represent scoriaceous or brecciated flow tops.

Structures resulting from tectonic deformation

Flattened and sheared pillow lavas. Deformation of pillow lavas is ubiquitous and the degree and nature of deformation varies with the position in the volcanic belt relative to shear zones and folds. Primary structures are completely obliterated in broad zones of intensely sheared amphibolite along the western margin of the formation and in most of the formation at the northwestern end of the belt. Passing inward into the belt from these zones, the first recognizable structures are intensely sheared, flattened and stretched pillows. The plane of flattening is invariably steeply dipping and trending roughly parallel to the length of the volcanic belt. Schistose foliation is subparallel to elongation of flattened pillows and limbs and axial planes of small-scale folds. In rare places, where shearing is least intense, fine penetrative crenulations distort rims of pillows and almost obliterate their structure. Weathered outcrops have elongate pock marks parallel to the foliation that in some places outline

relict pillows. Flattened fragments in deformed pillow breccias and rudaceous volcanic arenite vary from 2-20 cm long and 1-3 cm wide. Flattened pillows commonly have dimensions in the order of 20 x 100 cm.

Some round to oval pillows that appear undeformed in horizontal outcrop are highly elongate in vertical faces of outcrops. Elongate structures plunge steeply (about 80°) in the plane of tectonic foliation. Ratio of breadth to length of these bodies is commonly in the order of 1:10. Typical dimension of these oval pillows in a horizontal surface is 20 x 75 cm to 70 x 150 cm. In vertical exposures lengths are in the order of 300 cm.

Pseudo-layering. In many places deformed pillow lavas, pillow breccias or coarse clastic rocks are so intensely flattened and sheared that they have a layered appearance that resembles bedding. Pseudo-layering of this nature is found adjacent to large dykes, along fault zones, where clastic successions are deformed around rhyolite bodies, for example, at the south end of Sunset Lake, where rhyolite was apparently more competent than the mafic rocks during deformation, and within tightly folded successions. A tectonic origin of the pseudo-layering is indicated in some cases by (1) degradation of the layered aspect as less deformed rocks are approached, and (2) orientation of layering parallel to faults that cut across the general structural trend.

In places crudely layered rocks formed by tectonic processes may be impossible to distinguish from bedded units. In a ridge down the centre of the southern peninsula of Sunset Lake, for example, where pillow lavas and pillow breccias are intensely flattened, stretched and sheared, some layers, 5-6 m wide, with flaggy cleavage in the schistosity could be either local finer clastic units within the pillow succession or shear zones within the succession.

Locally, a layered appearance is produced where swarms of thin quartz veins have formed along finely cleaved amphibolite schists to produce alternating 20-40 cm thick layers.

Both relict primary layering and pseudo-layering are steeply dipping and parallel to the direction of flattening of pillow lavas and schistosity in the amphibolites. Both are locally deformed into parasitic folds on the limbs of larger isoclinal folds.

Lithology

The Sunset Lake Basalt comprises four main lithological variations: (1) medium to dark green amphibolite; (2) very dark green (almost black), very fine grained amphibolites; (3) pale green amphibolite; (4) layered amphibolites that are metamorphosed mafic volcanic sediments; and intensely sheared equivalents. Iron-formation forms minor units near the western margin of the formation.

Medium to dark green amphibolite. Most metabasalts (as in the Cameron River and Tumpline Lake subareas) are medium to dark green weathering, grey-green to green, fine- to medium-grained amphibolites. Locally, they

weather rusty brown adjacent to amphibolite dykes. In general, however, they are plagioclase and (?)pyroxene, microporphyritic metabasalts that range from aphanitic to fine grained and diabasic. Variolitic textures were noted at one locality where they form pale green, rounded to amoeboid bodies 1-2 cm across in massive basalt. Amygdules are more common, however, and are generally less than 2 mm across.

The Sunset Lake Basalt generally appear to be of lower metamorphic grade than rocks in the Tumpline Lake and Cameron River subareas. Consequently, their primary textures or relics are well preserved on a microscopic scale.

In thin section, relict recognizable primary textures include: amygdules, microphenocrysts of plagioclase in mafic minerals, feldspar microlites, and rarely, crystallites and spherulites. Amygdules, generally less than 2 cm across, are spheres to elongated ellipsoids that have smooth boundaries, except for local irregularities. Fillings, chiefly of quartz, calcite and epidote, have crude concentric zoning. Most commonly, very fine microgranular quartz (0.01-0.1 mm across) forms discontinuous rims and outer margins of amygdules, whereas coarser (0.02-0.3 mm) aggregates of sutured quartz grains form inner zones. In some cases, epidote forms a narrow zone between a microgranular rim of quartz and a core of coarse calcite.

Relict microphenocrysts of mafic minerals are interpreted from clusters of hornblende, actinolite or chlorite, that have sharply defined outlines that are coarser than relict microlitic texture of the matrix. The dominance of crude rectangular and short prismatic forms suggests that the original minerals were pyroxene.

Some specimens have remarkably well preserved seriate, porphyritic or glomeroporphyritic plagioclase in a matrix of felted to trachytic plagioclase microclites. Zoned plagioclase phenocrysts have broad cores and normal oscillatory zoned margins. Compositions range from An₃₀ to An₆₇. In more intensely recrystallized rocks microlites and microphenocrysts of plagioclase are preserved as polygonized laths.

Near locality S2 (Fig. 21), relict spherulites and crystallites are clearly preserved. Spherulites, ranging from 2-5 mm across, comprise minute acicular patches (5-7 microns wide and 100-250 microns long) of alternating chlorite and feldspar that form radiating fans. Each acicular area has minutely irregular boundaries and some areas coalesce to form elongate patches up to 30 microns wide that have the same optical orientation. A relict texture resembling feldspar crystallites in a glassy matrix is suggested by acicular aggregates (10 microns wide and 30-60 microns long) of microscopically minute feldspar surrounded by a matrix essentially of cryptocrystalline chlorite. This appears to be relict glass in which the spherulites formed.

Pale green amphibolite. A band of light coloured pillow lavas about 1 km wide lies along the western side of the Sunset Lake Basalt east of Sunset Lake (locality S1, Fig. 21). Outcrops weather pale green, pale grey to buff, and the fresh rock is pale green-grey and aphanitic. These basalts,

because of their pale colour and aphanitic texture, are easily mistaken for dacites in the field. Many field geologists have called them "dacite pillow lavas". Generally, they have high specific gravity, in the range of 2.80-2.95, except for a few localities where specific gravity is low (2.73-2.76). Pale green pillowed units are almost invariably basalts (chemically). Rocks that have abnormally low specific gravity are generally the most altered and contain abundant calcite. These rocks are essentially the same as darker pillow lavas farther east and in other parts of the belt, except that they are in the greenschist facies of metamorphism. Typical mineral assemblages are epidote-chlorite-tremolite-quartz-feldspar-calcite.

Microscopic primary textures are well preserved. Most rocks are sparsely plagioclase microphyric and have feldt to trachytic microlitic matrices. Microphenocrysts of plagioclase generally make up 1-2% of the rocks, and rarely, 30%. Plagioclase phenocrysts, having composition in the range of andesine, have weak, normal zoning at margins. Phenocrysts are generally less than 0.5 mm long, whereas microlites are slender laths less than 0.15 mm long. Microlites have two- to three-unit polysynthetic twinning.

Although plagioclase is remarkably fresh in most rocks, mafic minerals are invariably altered to chlorite and tremolite. Chlorite, tremolite and epidote fill areas between phenocrysts and microlites or replace mafic phenocrysts. Pseudomorphs of mafic phenocrysts (less than 1% of most rocks) have stubby rectangular prismatic forms and may be relics of pyroxene.

Quartz is present as amygdale fillings or in minor amounts in very fine granoblastic microfelsic aggregates between plagioclase and microlites.

Very dark green amphibolite. Very dark green metabasalts form most of the formation west of the Amacher Granite, the western 500 m of the formation along the eastern side of this pluton, and the eastern 1000 m of the formation due west of Payne Lake. These basalts are very fine grained to aphanitic amphibolites that weather very dark green to black and have exceptionally smooth polished glacial surfaces. They are dominantly pillow lavas and pillow breccias but are locally massive lavas. Pillows are generally very tightly packed and their thin rims are barely recognizable in most areas. Rare clastic material occurs in local pods. In one place the basalts form lava toes (mattress-shaped bodies) rather than pillows, that pass into massive lava, of a medium grained amphibolite in contrast to the fine grained amphibolite more typical of these rocks.

These amphibolites are the most mafic metabasalts in the formation. They contain 60-65% hornblende, have normative colour indices higher than most basalts of this belt, and high specific gravities (greater than 3.00), in the same range as amphibolite dykes.

Most rocks, even though they are highly recrystallized, contain some vestige of primary igneous minerals and textures. Aphanitic to very fine grained amphibolites tend to be recrystallized to a greater degree than fine grained metabasalts. Plagioclase laths are commonly polygonized;

more so in margins than in cores. Primary crystals (not polygonized) are normally zoned, have broad core (An_{60-70}) and thin rims (An_{39-40}). Metamorphic plagioclase displays itself as polygonized margins or complete polygonization of primary laths and as very fine irregular crystalloblastic aggregates (along with quartz) forming areas between coarser crystals.

The coarser grained rocks have relict subophitic texture where plagioclase laths are intergrown with hornblende. Blue-green crystalloblastic hornblende is interpreted to be recrystallized pyroxene in the original rocks. Relics of mafic phenocrysts are stubby forms up to 3 mm long that are completely replaced by chlorite and by some amphibole. Sharp euhedral outlines of the original crystals are well preserved. Original rocks were sparsely porphyritic basalts, having phenocrysts of plagioclase and probably pyroxene.

Volcanic sediments. A 100 to 200 m thick unit of thin- to medium-bedded (2-35 cm thick) volcanoclastic sediments (near locality S3, Fig. 21) that is traced for at least 2500 m, lies between very dark grey mafic basalts to the west and medium green weathering, less mafic basalts to the east. One graded bed suggests that the unit is overturned 75° to the west. The total extent of this unit was not mapped. Bedding is defined by layers of contrasting grain size (both primary clasts and coarseness of the amphibolite), colour and relief of the weathered surface. Some beds (1-2 cm thick) are internally finely laminated (0.5-2 mm). Beds containing finely disseminated pyrite weather buff grey in contrast to medium and dark grey weathering beds. Fine grained, thin-layered and laminated rocks are even bedded. Sediments are very dark green (almost black), resembling deformed arenites and shales, except that they are hard, very fine grained to aphanitic amphibolites. These rocks lack abundant mica in contrast to greywackes of the Burwash Formation. Arenaceous rocks contain sand to granule sized clasts of metabasalt in a very fine grained schistose amphibolite matrix. Clasts are flattened and elongated in planes parallel to bedding, cleavage and schistosity. This unit represents epiclastic rocks derived from basalts in the immediate area.

Conglomerate

Polymictic conglomerate forms a unit 1400 m long and up to 30 m wide within the western margin of the Sunset Lake Basalt, about 500 m west of Amacher Lake (Fig. 22). Although the unit is in contact with pillow lavas of the Sunset Lake Basalt in only one small area, and makes faulted contact at one locality with the western side of one of the granitoid lenses in the volcanic belt, for the most part it is surrounded by a complex of mafic dykes. Near the central part a thin dyke splits the conglomerate longitudinally. Where the conglomerate is in contact with granitoid gneiss, the latter is cataclastically deformed. This deformation and the presence of mafic dykes obscures the original contact relationships.

This unit is a clast supported polymictic conglomerate comprising pebbles, cobbles and boulders, in decreasing order of abundance, of fine grained metabasalt, medium- to coarse-grained metagabbro, vein quartz and foliated grani-

toid gneiss. Clasts, which are flattened and elongated in a strongly foliated mafic-rich matrix (Fig. 23), have variably oblate shapes with aspect ratios ranging from 6:6:1 to 12:6:1. Clasts are subangular (largest are volcanic and gabbro fragments) to rounded (quartz and granitoid fragments) and have maximum sizes as follows: volcanic, 30 × 60 cm; metadiabase and gabbro, 20 × 40 cm; vein quartz, 4 × 10 cm; and granitoid, less than 10 cm across. Largest clasts tend to be concentrated in crude layers and locally the outcrop shows a rough grading toward the east. In most places this is essentially a volcanic-gabbro conglomerate containing sparse (1-2%) quartz and granitoid clasts.

The conglomerate is derived mainly from the enclosing volcanic pile and its related intrusives. The foliated granitoid clasts, however, suggest proximity to a granitic terrane that had a predepositional history of deformation (i.e. possibly the Sleepy Dragon Complex). The abundance of mafic dykes and rare exposures of the faulted contact make it unclear whether the conglomerate-granitoid contact is a faulted unconformity or completely tectonic. The fault-bound nature of the granitoid lenses, however, support the latter.

Iron-formation

West and southwest of Amacher Lake banded iron-formation has been traced intermittently for 13 km, where it occurs at the boundary between sheared granitoid gneiss and volcanics or within amphibolite dykes. Although it is now several separated units, ranging from 1-50 m thick, the original form is obscured by the large intrusions. Where the iron-formation has been mapped in detail, west of Amacher Lake (Fig. 22), much of it is totally enclosed within an amphibolite dyke complex. At one locality a thin sliver of iron-formation is in contact with volcanic conglomerate. The iron-formation which is up to 22 m wide, is folded and perhaps split apart locally within the dyke complex. It may not be as continuous, therefore, as inferred in Figure 22. Contacts with surrounding amphibolite are both conformable and discordant with respect to compositional layering. In places the abrupt termination of the iron-formation may be due to faulting or attenuation.

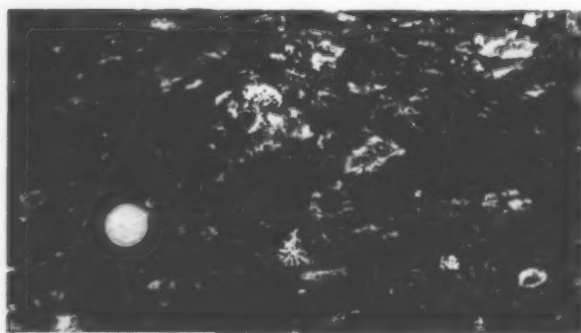


Figure 23. Polymictic conglomerate in Sunset Lake Basalt west of Amacher Lake. GSC Z03714-K

Because of its well developed millimetre- to centimetre-scale compositional layering, the iron-formation provides some of the best evidence of polyphase deformational history within the volcanic belt (Fig. 24). This layering is parallel to a strong, locally mylonitic, foliation in the surrounding volcanics and granitoid gneisses or schists. Overprinting relations demonstrate at least 3 generations of deformation by folding (Lambert and van Staal, 1987). Amplitudes of folds are generally less than 2 m and most minor folds have amplitudes in the order of 10 cm. Thus a variety of complex structures reflecting this high strain and fold interference include tight to isoclinal, asymmetrical structures, boudinaged rootless intrafolial folds, sheath folds, hook and mushroom shaped structures (types 2 and 3 of Ramsay, 1967; see Structural geology).

The rock is a recrystallized chert magnetite iron-formation (Fig. 25). It comprises alternating white to grey layers (less than 1 cm thick) and black to dark-green mafic layers (0.5-6 cm thick; averaging 1-3 cm) containing assemblages: (1) quartz and minor magnetite, (2) dominantly amphibole and magnetite, and (3) dominantly magnetite and minor quartz and amphibole. Granoblastic aggregates of quartz are interpreted to be recrystallized chert. Coarseness of the aggregate varies from layer to layer and the boundary between layers is sharp. In general, layers of coarse granoblastic quartz (grains 0.05-1.0 mm) contain very minor magnetite and hornblende; layers of finest granoblastic quartz (0.02-0.05 mm) contain abundant, very finely disseminated magnetite; and layers rich in coarse magnetite contain moderately coarse, granoblastic quartz. Amphibole is pale blue-green actinolitic hornblende that has strong preferred orientation parallel to layering. It is most abundant in magnetite-rich layers. Epidote occurs in minor amounts, generally associated with amphibole.

The iron-formation is interpreted as a single unit that has been folded, faulted and split apart by a complex mafic intrusion. Some deformation may have taken place during injection of the intrusions. The iron-formation may represent a local eruptive quiescence in the voluminous effusion of mafic lava that formed the Sunset Lake Basalt. It may also



Figure 24. Doubly plunging F_2 folds in polydeformed iron-formation near Amacher Lake. GSC 170811.

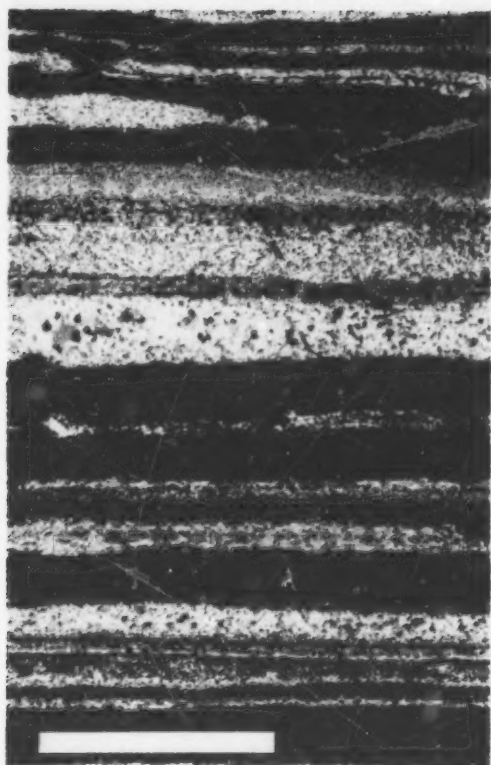


Figure 25. Photomicrograph showing chert-magnetite iron-formation near Amacher Lake. Plane light. Scale bar 1 mm. GSC 190983

have been a plane of weakness in the volcanic pile which was the locus of dyke intrusion.

Interpretation

The Sunset Lake Basalt represent voluminous outpouring of magma in a subaqueous environment. The dense swarms of amphibolite dykes near the sides of the belt are assumed to be the subplutonic equivalents of the pillow lavas and to mark positions of eruptive centres for the basalts. That the earliest effusions were the most mafic lava is suggested by the distribution of the very dark green basalts along the western and eastern sides of the belt. Later lavas within the formation are only slightly more felsic basalts. Apparently the composition of magma became more felsic with time, as indicated by the passage of the Sunset Lake Basalt into andesites of the overlying Alice Formation. No major break in eruptive activity nor any period of erosion is indicated in the stratigraphy. That the Sunset Lake Basalt was still erupting when the Alice Formation andesites were being deposited, is suggested by the interfingering relationships between the two units northwest of Sunset Lake. Volcanic sediments within the volcanic belt, iron-formation and conglomerate may represent a period of local eruptive quiescence during which clastic debris derived from the pillow lavas and hyaloclastites and chemical sediments were

deposited. Foliated granitoid boulders in the conglomerate document the presence somewhere of a deformed granitoid terrane.

Alice Formation

Definition, distribution and thickness

The Alice Formation is a new name that applies to units of andesitic to dacitic rocks that lie in a northerly trending belt centred on Sunset Lake (Fig. 26). The formation is named after the Alice Claim Group, on the east side of Sunset Lake, where the Sunset Yellowknife Mines Ltd. developed a gold property between 1945 and 1947 (Lord, 1951, p. 273-276).

The formation is divided into two members: an andesite member which includes pillow lavas, breccias, massive flows and minor pyroclastics and volcanic sediments; and a dacite member including felsic andesite to dacite tuffs, volcanic breccias, lavas and cataclastic equivalents. The formation appears to occupy the centre of a broad synform that plunges gently south and ends abruptly near the south end of Sunset Lake. In this region the formation interfingers with Sunset Lake Basalt and rhyolites, or ends abruptly against the Burwash Formation. The top of the dacite member is not exposed in most areas so the maximum thickness of the formation is not known but may be in the order of 1200 m. The andesite member may be up to 800 m thick.

Contact relations

In most places the formation overlies conformably or interfingers with the Sunset Lake Basalt. Near the southwestern end of Sunset Lake fingers of the formation swing southwesterly and tend to conform with the contact between the volcanic belt and Burwash Formation. It is not clear how much of this interfingering is due to interlayering of primary units, repetition by folding, or deformed lobate flows.

Near the southeast end of Sunset Lake some units swing southeasterly. On the peninsula at the south end of Sunset Lake, however, foliation in the dacite member trends at a high angle to the volcanic-sediment contact, suggesting that a local unconformable relationship may exist between this member and the Burwash Formation.

At one locality southeast of Amacher Lake, thin beds of volcanoclastic sediments mark the contact between Sunset Lake Basalt and the Alice Formation. In general, however, there is no distinct marker horizon between the andesite member and the Sunset Lake Basalt. Contacts are based on change in lithology within the thick succession of pillow lavas. Southwest of Sunset Lake the lithological contrast is obvious and contacts are drawn with certainty. West and north of Sunset Lake, however, lithological contacts between andesite and basaltic pillow lavas are commonly not easily distinguished. Hence the contact on the map is shown as an approximate boundary that marks the vicinity in which there is a transition from andesitic to basaltic compositions.

North of Sunset Lake, bodies of rhyolite both cut across the Alice Formation and form layers within the formation.

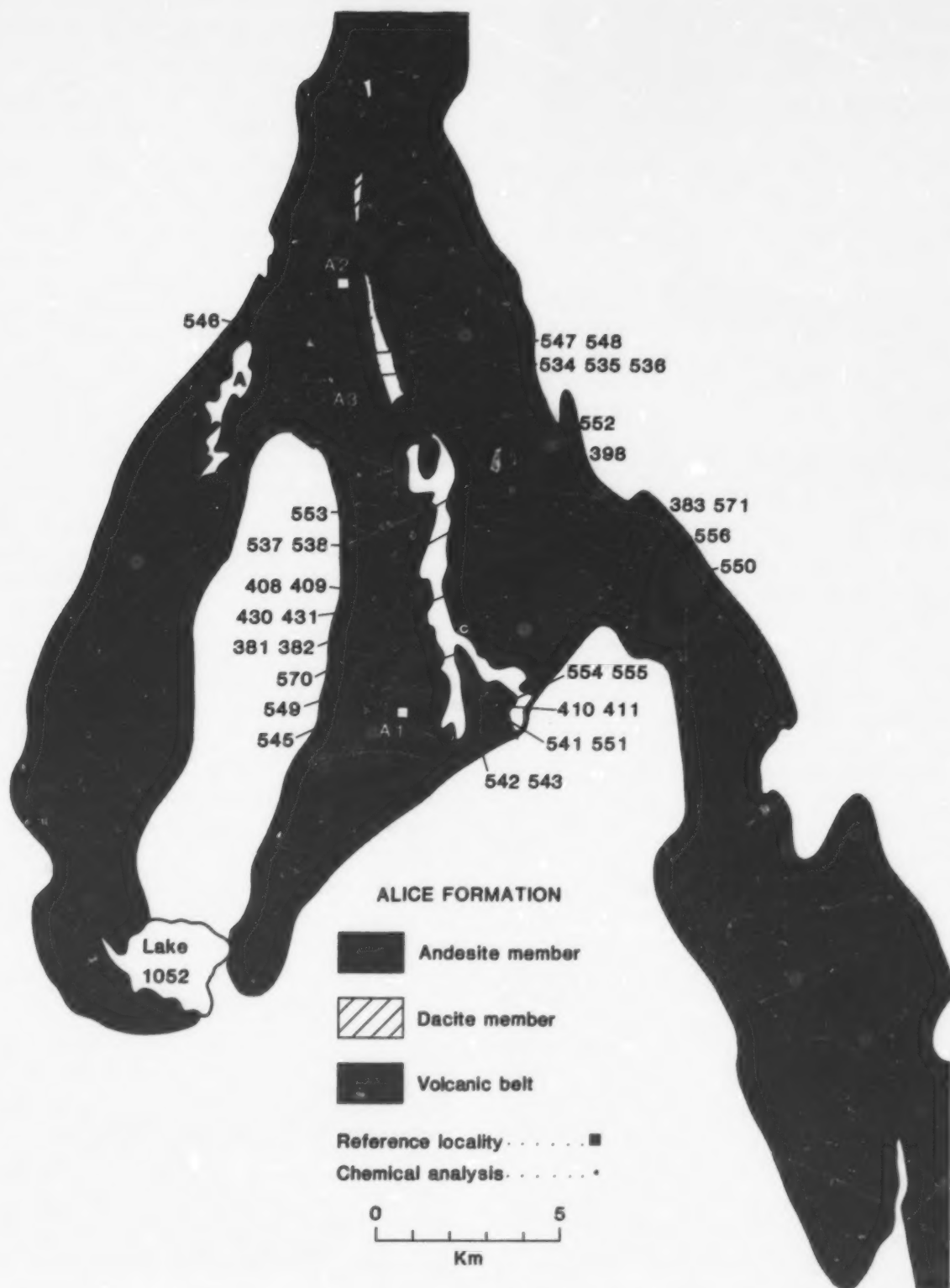


Figure 26. Distribution of the Alice Formation; reference and chemical analysis localities. A, Amacher Lake.

The boundary between the andesite and dacite members is shown as an approximate or assumed contact in Figure 3. This boundary corresponds in most places to the specific gravity contour of 2.75 for rocks in this formation. Rocks of the dacite member have specific gravity less than 2.75 (2.67-2.75) whereas those of the andesite member have specific gravity ranging from 2.75-2.85. Generally, the former have lower normative colour index and normative plagioclase compositions than the latter and are chemically dacites. In most places, however, one or more of the following lithological and outcrop characteristics change when passing from the andesite member to the dacite member. (1) Colour of weathered outcrop becomes paler; generally it changes from grey green or medium grey or green (andesite member) to pale grey, pale green or buff grey. (2) The colour of the fresh rock becomes paler due to lower content of mafic minerals. (3) There is a change from pillow lavas and pillow breccias (andesite member) to massive, tuffaceous and cataclastic rocks of the dacite member. (4) Massive and tuffaceous units (predominant in the dacite member but also present in the andesite member) are cataclastically deformed and have well developed slaty cleavage, or are phyllites. (5) In some places there is an abrupt change in slope where dominantly pillowed units give way to recessive weathering clastic units. North of Sunset Lake the western boundary of the dacite member is marked by a thin unit of rhyolite that appears at intervals for about 3 km. A smaller rhyolite unit marks this boundary about 1 km from the southwestern side of Sunset Lake.

Stratigraphy and lithology

Andesite member. The andesite member comprises closely packed pillow lavas with local areas of clastic rocks that show every gradation from broken pillow breccias and isolated breccias to fine hyaloclastites. Massive lavas are uncommon except near locality A1 (Fig. 26, southwest of

Sunset Lake). Near this locality the andesite member contains the following succession from west to east. The lower 500 m is pillow lavas of which the upper (eastern) 50 m are amygdaloidal. Pillow lavas are overlain by 250 m of massive andesitic lavas. These lavas are buff to grey weathering, dark grey to green (fresh) andesites that have local vesicular and amygdaloidal zones. Lavas pass eastward into 100 m of felsic fragment lapilli and block breccias. All clastic rocks are strongly flattened and stretched.

Lavas contain sparse to abundant (up to 30 %) phenocrysts of plagioclase in a medium grey to green-grey aphanitic groundmass comprising plagioclase microlites (Fig. 27) and, commonly, abundant epidote interstitial to the microlites. Epidote also occurs as euhedral grains and clusters of grains (up to 250 microns) that have outlines resembling phenocrysts of pyroxene, and probably are pseudomorphs of pyroxene. Andesine phenocrysts are euhedral to resorbed, and generally unzoned. Abundantly porphyritic rocks have a seriate porphyritic or glomeroporphyritic texture. These rocks generally have well preserved plagioclase microlites that have strong preferred orientation and stream around phenocrysts, defining a flow texture. Much of the andesite pillow lava within 1 km of the southwestern side of Sunset Lake weathers pale green to buff and is pale green on fresh surface. This pale colouration is in part due to abundant microscopic calcite in some areas, and to abundant microscopic epidote in other areas. Broken pillow breccias at locality A2 (Fig. 26) comprise about 2 % grey-green blocks up to 20 × 40 cm, in a medium grey-green matrix of nonsorted volcanic arenite (hyaloclastite). Most blocks are rounded and some have distinct pillow rims that weather in relief above the core. Coarse blocks occur along a particular horizon which may indicate crude layering. The volcanic arenite is generally massive.

Microscopic detail of hyaloclastite is rarely well enough preserved to be clearly recognizable. In one area,

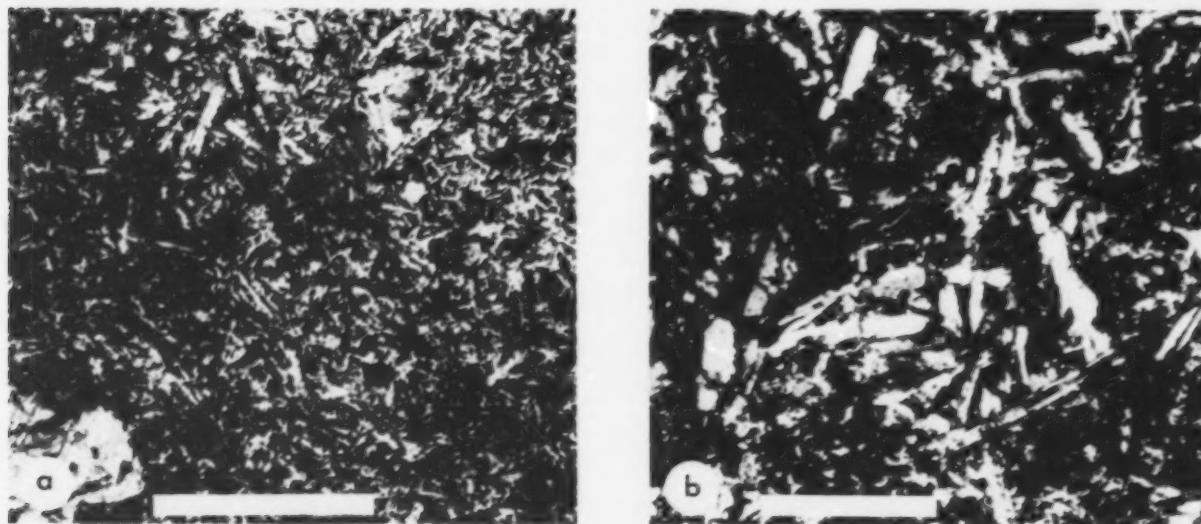


Figure 27. Microlitic textures in andesite lavas of the Alice Formation west of Sunset Lake. (a) scale bar 1 mm (GSC 190908), (b) scale bar 0.2 mm (GSC 190953).

hyaloclastite within the pillow breccia unit is essentially a recrystallized glassy rock. Relics of shards, preserved as irregular granoblastic felsic aggregates, are in a matrix of chlorite, epidote and minor quartz-feldspar aggregate. Forms of shards vary from smooth-sided, rounded to arcuate blebs, angular triangular shapes (rare), blocky rectangles and cuspidate to rounded blebs; vesicles are rare.

Pyroclastic rocks are rare in the andesite member. At chemical analysis locality 546 (Fig. 26) a unit is interpreted as scoriaceous block tuff-breccia. The rock comprises about 30% ovoid to lensoid fragments up to 15 cm long and 6 m wide that have strong preferred orientation in a brownish-grey weathering clastic matrix. Pale grey to green weathering fragments contain abundant round to irregular pock marks that are not flattened, and are reminiscent of scoriaceous lava or pumice. Shapes range from smooth-sided lenses, angular, elongate to equant blocks to ovoids that have irregular, ragged boundaries. The matrix contains smaller angular, elongate to platy fragments and irregular elongate wisps (resembling scoriaceous glass) that tend to parallel fragments or stream around them.

The deformed nature of these rocks is most evident in clastic units. Locally, large blocks (pillows) form pinch-and-swell lenses that have preferred orientation in a gritty looking matrix. Clasts that appear undeformed on horizontal surface generally show some degree of steeply plunging elongation in vertical surfaces. Tuffs commonly have highly elongate mafic and felsic streaks (on a millimetre scale) that are equant in cross-section. Thus, deformation is greater than it looks on a macroscopic scale. Generally, flattened bodies have a preferred orientation with a northerly trend almost parallel to the valley of the Beaulieu River and the trend of Sunset Lake. Shearing in some outcrops is so intense that it obliterates pillows. Such outcrops and massive outcrops commonly have a flaggy cleavage parallel to the shear plane that weathers out as deep pock marks.

Some macroscopically massive lavas are protomylonites that are microscopically a mass of minute anastomosing shears that define a subtle foliation. The weathered surface of these rocks is rough and pitted and resembles deceptively fine grained clastic rocks or tuffs. Evidence of tectonic deformation is (1) undulatory microshears that divide rocks into closely packed lenticles (Fig. 28), (2) broken and boudinaged phenocrysts, and (3) recrystallization trails that taper off from the ends of phenocrysts in the plane of foliation. Protomylonites of the andesite member (northwest of Sunset Lake) are similar to those in the dacite member, except that they lack abundant rhyolite fragments, have higher specific gravity, and are generally darker.

Dacite member. The dacite member of the Alice Formation includes lavas (both massive and rarely pillowed), tuffs, breccias and cataclastic equivalents of these rocks as well as slates and phyllites. Most of these rocks have the chemical composition of dacites, but some of the pillowed units are felsic andesite. They are buff, pale grey, pale green weathering rocks that are medium grey and aphanitic on fresh surface. Some felsic lavas contain abundant fine calcite which has had the effect of increasing the normative

plagioclase composition and causing the rock to appear chemically as an andesite.

Autobrecciated massive lavas (A3, Fig. 26) comprise large to small angular blocks of grey dacite having areas between blocks filled with finer fragments and grit of the same material.

Felsic andesite pillow lavas that lie between the two rhyolite units at locality A3 (Fig. 26) are pale green weathering rocks containing 10-40% phenocrysts of plagioclase in an aphanitic matrix. These seriate porphyritic rocks (Fig. 29) contain euhedral to rounded phenocrysts of plagioclase, ranging from 0.5 to 5 mm long, that occur as single crystals or glomeroporphyritic clusters that have a marked preferred orientation. Crystals are not zoned and range in composition from albite to oligoclase (An_{2-11}). Matrix is a mass of microlites and skeletal crystals (0.01-0.03 mm) that have random to strong preferred orientation where they stream around phenocrysts and quartz amygdules. Some rocks have abundant quartz amygdules, now spherules, 1-5 mm across of polygonal quartz.

Tuffs and pyroclastic breccias are preserved near the southwestern side of Sunset Lake, along the east side of the

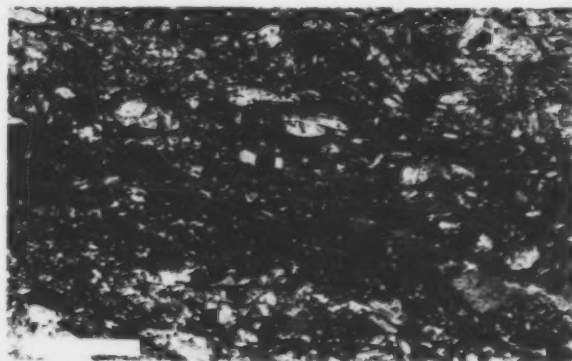


Figure 28. Lenticular protomylonite in Alice Formation andesite west of Sunset Lake. Scale bar 2 mm. GSC 190981

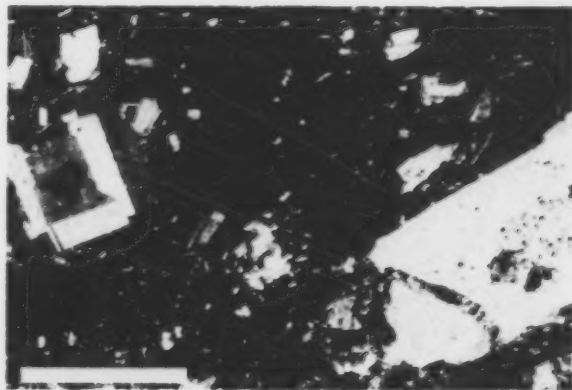


Figure 29. Seriate feldspar-phyrlic felsic andesite of the Alice Formation north of Sunset Lake. Scale bar 1 mm. GSC 190928

southern peninsula of Sunset Lake, and west of the Beaulieu River about 2 km north of Sunset Lake. Near the southwestern side of Sunset Lake, lapilli tuff forms a lens about 170 m thick within the andesite member. This unit overlies about 300 m of massive andesitic lavas and underlies a thin tongue of pillowed basalts. This massive tuff unit contains fragments up to 4 cm long of white amygdaloidal rhyolite pumice and shreds of pale green-grey felsite. Some clasts are quartz-phyric rhyolite that have irregular undulatory and wispy margins. Matrix to the tuff is a dark grey coarse ash tuff comprising equant to elongate shapes (possibly relict shards) containing abundant microphenocrysts of feldspar.

Dacite-tuff, -lapilli tuff and -block tuff north of Sunset Lake overlies and underlies the easternmost unit of rhyolite. Pyroclastic units are in the order of about 50 to 100 m thick. Some of the dacite overlying the uppermost clastic units is plagioclase-phyric massive lava. A minor amount of pillow lavas occurs in one place above the tuff. Both pyroclastic units are similar in that they contain abundant amygdules and elongate fragments (maximum size ranges from 2-8 cm) of rhyolite, phenocrysts and crystal fragments of quartz and feldspar, fragments of porphyritic dacite, and aphanitic felsite in a pale green matrix. The matrix is grey to green, chloritic grit or aphanitic material that has a chloritic sheen on fine undulatory cleavage surfaces which permeate the rock. Weathering along these surfaces gives outcrops a pitted appearance. The rocks are poorly sorted and nonbedded.

Along the southeastern side of the southern peninsula of Sunset Lake, andesitic pyroclastic rocks show a good lenticular texture in some outcrops that resemble welded tuffs in modern rocks. Some lenticles weather with a pitted texture resembling pumice. These rocks are recrystallized tuff containing phenocrysts and crystal fragments of euhedral to subhedral plagioclase and quartz and clasts of microporphyritic and microlitic andesite up to 2 cm in the matrix comprising fragments of quartz and plagioclase crystals and crystalloblastic felsic material in which relict shards are rarely recognizable. Boundaries of microlitic clasts vary from irregular, angular, cuspidate to smooth, and some resemble shards. Blue-green hornblende forms poikiloblastic laths or radiating patches (that have grown in bowtie-like patterns) and have grown across the fabric of the matrix. These rocks are void of textures resembling those in lavas (except in some clasts). Specimens studied show no microscopic evidence of welding. The apparent eutaxitic foliation in outcrops may be the result of tectonic deformation or of growth of mafic minerals in patches along tectonic planes.

Most rocks of the dacite member show some degree of brittle and ductile deformation. Rocks commonly have well developed flaggy to slaty cleavage that is parallel to the major northerly trending lineaments defined by the valleys of the Beaulieu River and Sunset Lake. The finest grained rocks are slates and phyllites, whereas massive lavas and some tuffs and breccias are cataclases and protomylonites (as defined by Sibson, 1977).

Pale green to grey slates and phyllites occurring along the eastern side of Sunset Lake and locally along the Beaulieu River north of Sunset Lake are soft aphanitic rocks that have no distinctive clastic texture. They contain crystals

of euhedral to anhedral plagioclase (about 0.3 mm long) that vary from fresh to almost completely saussuritized in a crystalloblastic matrix of strongly oriented quartz and feldspar with chlorite, and minor calcite. These rocks are interpreted as metamorphosed and deformed, very fine grained volcanic clastic rocks that may have been tuffs or pelitic sediments.

Protomylonitic rocks form massive outcrops that have finely pitted weathering surfaces. In many places, cataclastically deformed lavas are almost indistinguishable megascopically from massive tuffs (this is true of the western side of Sunset Lake). Identity of the original rock is generally revealed only by textures within microscopic lenticles that make up these rocks.

Cataclastic deformation appears in various degrees from rocks that are broken by anastomosing networks of cleavage through microbreccias to crushed lenticular rocks grading into protomylonites (Fig. 28). Most intensely deformed mylonitic rocks are a myriad of microscopic lenticles, slivers and streaks of rock fragments and plagioclase crys-

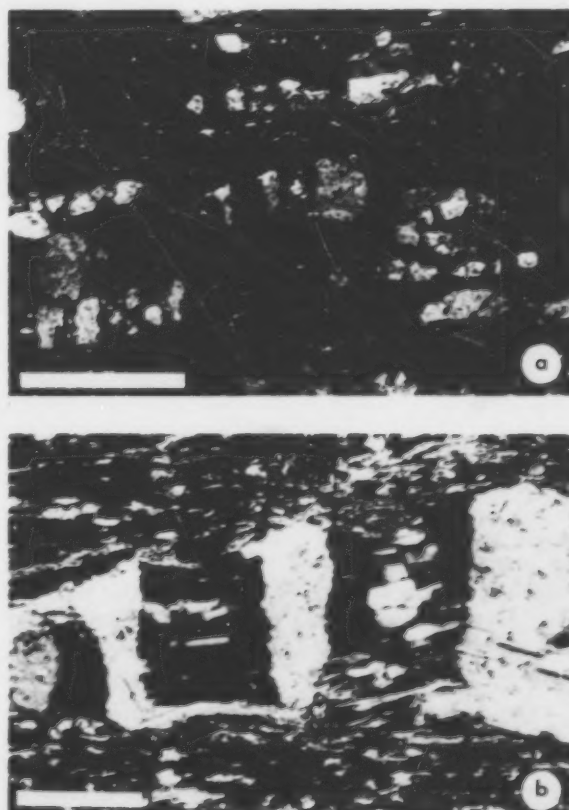


Figure 30. Photomicrographs of sheared rocks from the dacite member of the Alice Formation, northwest side of Sunset Lake. (a) Microscopic lenticles having strong preferred orientation: Scale bar 1 mm (GSC190964). (b) Fibrous crystallization between boudinaged parts of plagioclase phenocrysts. Scale bar 0.2 mm (GSC 190971).

tals that have strong preferred orientation in a recrystallized, finely comminuted matrix containing extremely fine sericite, biotite, chlorite and minute crystalloblastic quartz and feldspar (Fig. 30). Calcite and epidote are disseminated throughout the rock, replace rock fragments, and form veins parallel to foliation. Porphyritic and microlitic textures of lavas are preserved within the lenticles. This tectonic lenticulation and crushing has broken and fractured plagioclase phenocrysts. Matching aligned segments of boudinaged phenocrysts (Fig. 30b) are separated by a homeoblastic matrix of strongly oriented elongate crystals of quartz and feldspar. Plagioclase is partly to completely polygonized but commonly maintains its euhedral outlines in spite of the recrystallization and intense saussuritization. Relics of quartz phenocrysts are completely recrystallized to rounded aggregates of granoblastic crystals that have long crystallization trails tapering out in the plane of foliation.

Cataclastically deformed tuff, near chemical analysis location 398 (Fig. 26), is a microbreccia made up of fragments of lithic tuff containing plagioclase-phyric lava, plagioclase crystals, and relics of shards. Relict shards within the fragments are granoblastic felsic mosaics that have rounded, cuspidate, bubble wall and platy forms. The area between the fragments is a recrystallized, fine aggregate of tuff fragments that commonly contain abundant chlorite and calcite alteration.

Interpretation

This formation records a major change in the composition of volcanic products, type of eruption and environment of eruption within the Sunset Lake subarea. These changes are progressive with time and appear to have taken place near the waning stages of volcanism (suggested by the relative small amount of these rocks compared with the total exposed volume of the rocks in the volcanic belt). Compositions of eruptive products change from dominantly basaltic to andesites and dacites. The change from dominantly pillow lavas of Alice andesites to pyroclastic rocks of the Alice dacites indicates a change from quiet effusion of lava to explosive eruption. The association of pyroclastic rocks in both members, the massive lavas, and vesicular pillow lavas, suggest that eruption took place in a shallow water or possibly a subaerial environment. Finally, the interfingering of rhyolite units with pyroclastic rocks of the Alice dacites north of Sunset Lake near rhyolite to dacite domes suggests that the domes are consanguinous with parts of the dacite member and may mark vents of some of the pyroclastic eruptions.

These relationships suggest that the formation began with submarine effusion of andesite and felsic andesite pillow lavas which gave way to explosive eruptions during the latter stages of eruption when the volcanic pile had built up to a shallow water environment or possibly emerged locally above sea level.

Rhyolite

A series of rhyolite bodies in the Sunset Lake subarea lie at or near the top of the volcanic succession and commonly

form a unit between the Sunset Lake Basalt and the Burwash Formation (Fig. 31).

The rhyolites occur in three main areas: (1) a thick body at Lake 1052, (2) a dome at the south end of Sunset Lake, and (3) a dome complex near the north end of Sunset Lake. Rhyolite dykes and dyke swarms (ranging from 2-40 m thick) in the mafic volcanics lower in the volcanic succession, are considered to be consanguinous with the main rhyolite bodies. In the southwestern corner of the Sunset Lake subarea, the rhyolite correlates with the Turnback Rhyolite.

The rhyolites have a maximum apparent thickness (about 1100 m), west of Lake 1052 thin near the eastern side of the lake and appear to finger out between lake 1052 and Sunset Lake. Some of this apparent interfingering could be due to repetition by folding or rhyolite that filled isolated valleys in the original topography. The precise internal structure of the volcanic belt in this area is dubious. A northerly trending sinistral fault east of Lake 1052 has displaced the formation about 1 km.

The dome at the south end of Sunset Lake has a maximum thickness of about 1100 m but pinches out abruptly within 1000 m to the north, and more gradually 3500 m to the southwest. The dome complex north of Sunset Lake comprises three rhyolite units superimposed on one another and locally separated by units of andesite and dacite. The thickest rhyolite unit of the complex is about 500 m thick, and the complex as a whole may be about 1000 m thick. Aprons of pyroclastic and coarse epiclastic debris, generally less than 100 m thick, extend for 2 to 6 km to the north of the complex and thin gradually northward.

Contact relations

Rhyolite lies conformably between the Sunset Lake Basalt and the Burwash Formation near Lake 1052 and appears to interfinger with andesitic to basaltic volcanics southwest of Sunset Lake. Near the south and east sides of Lake 1052 layered felsic volcanics and rhyolite breccias make conformable contact with metasediments of the Burwash Formation. At the upper contact near the southern and eastern sides of the lake, numerous layers of felsic volcanics occur within the Burwash Formation. Southwest of Lake 1052 units of carbonate and mafic volcanics (less than 85 m wide) lie between the rhyolites and the Burwash Formation.

The dome at the south end of Sunset Lake lies conformably on mafic to intermediate volcanic rocks. Units of mafic volcanics appear to lap up on the central (eastern) sides of the dome where they taper out to the north and to the south. Mafic and felsic volcanics are interbedded near the upper contact of the dome where it tapers southerly. Segments of this dome are displaced by easterly trending faults, so that parts of it make fault contact with the felsic andesites.

Units of the dome complex north of Sunset Lake have conformable lower contacts, whereas upper contacts interfinger with onlapping units on the southern side. Tuff and breccia units extending north of the dome lie conformably within andesitic and dacitic tuffs of the Alice Formation.

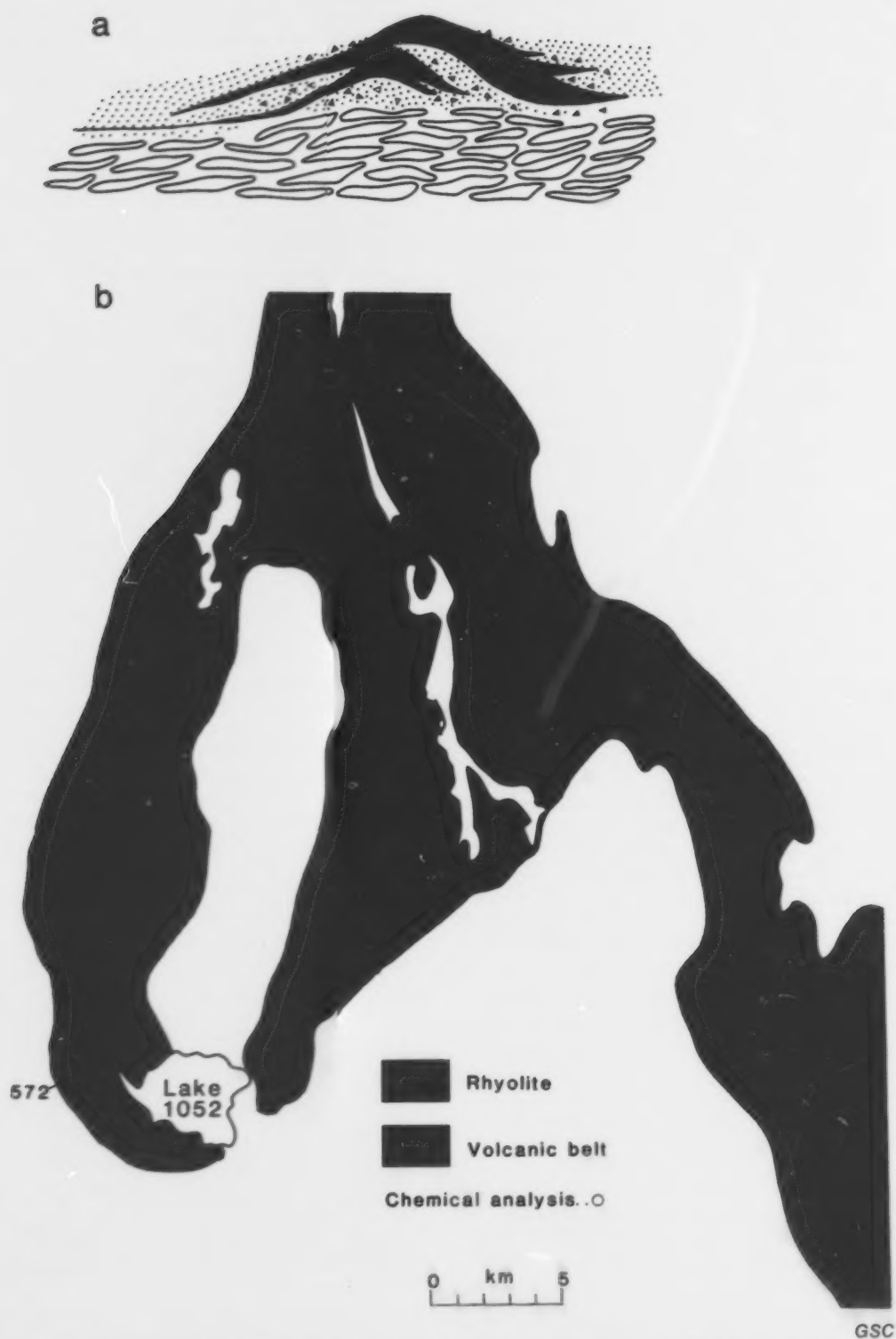


Figure 31. (a) Palinspastic reconstruction of rhyolite complex (black) north of Sunset Lake. Rhyolite and associated breccias and tuff overlie pillow lavas of the Alice Formation. (b) Index map showing rhyolites (black) in the Sunset Lake subarea; reference and chemical analysis localities.

Rhyolite near Lake 1052

Massive rhyolite southwest of Lake 1052 varies from massive to layered rhyolite east of the lake, and is dominantly layered rhyolite east of the major fault where the unit tapers out. The formation is dominantly white to buff weathering, massive, porphyritic rhyolite in which euhedral to subhedral, rounded to embayed phenocrysts of quartz (1-2 mm across), that make up 2-10% of the rock, lie in a microfelsitic aggregate of quartz and feldspar that has granoblastic to very irregular crystalloblastic texture. Altered varieties in fault zones are greenish-yellow rocks that have a resinous lustre.

Southwest of Lake 1052 a pinching and swelling unit of mafic volcanics and lenses of carbonate occurs in a zone about 85 m wide at the upper contact of the rhyolite. In this zone the rock grades (over a distance of up to 60 m) from massive rhyolite to rhyolite breccia with carbonate matrix into dominantly carbonate-bearing rhyolite clasts. Near the contact with mafic volcanic units, the carbonate contains both mafic and rhyolite clasts.

In the vicinity of the northerly trending faults, between Lake 1052 and Sunset Lake, apparent layering and eutaxitic-like structures are tectonic features parallel to the faults, and not bedding. These structures include (1) prominent lenticular cataclastic structures, and (2) apparent layering formed by weathering along closely spaced planes of schistosity that define trains of pock marks or recessions (spaced 1-20 cm apart). In one place this tectonic foliation is folded into tight chevron folds having amplitudes of 1-3 m and axial planes about parallel to the trend of the main tectonic foliation of the outcrop.

Dome at south end of Sunset Lake

Much of the thickest part of this dome is massive to brecciated rhyolite. Some rocks in the dome are lenticular, resembling eutaxites, but lenticles have the same trend as the easterly trending faults and are thus considered to be cataclastic. Near the base of the thickest part of the dome, a lenticular foliation trends parallel to contacts of the body. This foliation could be primary or a tectonically flattened primary breccia. Rocks in the southern flanks and top of the dome are thin-bedded (0.5-2 cm) or flow layered.

At the thickest part of the dome rocks range from white weathering, pale grey metarhyolite on the top (eastern side) and on tapering flanks, to pale grey weathering, dark grey dacite and possibly andesite near the base (western side) at the thickest part of the dome. This change is marked by (1) specific gravity increasing from 2.65 to 2.69 in the rhyolite to 2.75 to 2.76 in dacites, (2) change in colour of fresh rock from pale- to medium-grey in rhyolite to dark grey and black in dacites. Some dacites have distinct lenticular structure oriented parallel to formational contacts, defined by mafic-rich and mafic-poor areas. Although such rocks are probably relict clastic units, no indubitable pyroclastic textures were recognized.

Locally, layering (0.5-2 cm wide) is marked by (1) variation in mafic content (alternating layers rich and poor in hornblende and biotite — up to 20%) and (2) variation in

the abundance and population of phenocrysts (felsic layers tend to have less than 1% microphenocrysts of quartz, whereas darker layers contain 1 to 5% phenocrysts of plagioclase, potassic feldspar and quartz up to 2 mm) and (3) texture of matrix (felsic layers have irregular crystalloblastic, very fine microfelsic matrix, whereas phenocryst-rich "mafic" layers have abundant plagioclase microlites that have strong preferred orientation parallel to layering and plagioclase phenocrysts). In darker layers plagioclase phenocrysts are more abundant than quartz phenocrysts. In some rocks, millimetre-size phenocrysts are completely polygonized, but interpretable as plagioclase from their relict outlines. Relics of plagioclase microlites are almost completely polygonized but are recognizable by lath forms and twin units that have dominantly the same extinction. The rocks may contain up to 20% calcite, some of which has completely replaced plagioclase phenocrysts; some is irregularly dispersed throughout the rock; and some forms amygdule fillings. Layering, porphyritic (whole crystals rather than crystal fragments) and microtrachytic or pilotaxitic textures suggest that the dacites are lavas.

This body is interpreted as an exogenous dome that has shed aprons of debris towards the south. This dome may be the eruptive centre of felsic andesite to dacitic pyroclastic rocks that lie to the north.

Dome complex north of Sunset Lake

The dome complex, centred at the north end of Sunset Lake, comprises overlapping bodies partly separated by intervening units of intermediate volcanics. The complex is fragmented by the north-northeast-trending sinistral strike-slip fault and by a swarm of northerly trending faults splaying from the major fault. Figure 31a shows a palinspastic reconstruction of the rhyolite complex. The upper (southeastern) and largest body of the complex is dominantly massive, white weathering, quartz-plagioclase-phyric (and locally, potassic feldspar) rhyolite. It contains areas of coarse brecciation near the top and on the southeastern flanks (these are primary breccias not tectonic breccias). A unit of rhyolite boulder conglomerate occurs within the apical breccia zone. The lower body is massive, in part, at its apex, but on its flanks, is dominantly coarse lithic breccia, which passes northward through tuff breccias to lapilli tuffs and crystal tuffs. The westernmost thin unit, that trends northward from the dome complex, comprises clastic rocks including tuffs and rhyolite volcanoclastic sediments. Numerous rhyolite dykes cutting intermediate to mafic volcanic rocks to the west of these rhyolite bodies, have lithologies that suggest they are related to the complex.

Breccias and conglomerates. Megabreccias near the apex and upper flanks of the upper rhyolite dome contain angular to lensoid blocks of rhyolite 3-15 m across in a matrix of finer breccia. A rhyolite boulder conglomerate at locality R1 (Fig. 31) passes into rhyolite block breccias, which in turn grade into massive and flow layered rhyolite.

The conglomerate comprises 70 to 80% well rounded pebbles and boulders of quartz-phyric rhyolite (ranging from 0.5 to 25 cm across and averaging 10 cm) in a dark

grey grit matrix containing subangular rhyolite fragments, less than 5 mm across, in a mesostasis of carbonate (Fig. 32). Locally, the carbonate-rich matrix makes up 75 % of the rock. Pyrite is disseminated in the matrix and along fractures in clasts. Boulders have dark grey cores and white weathering rinds. They are elongate and have northerly orientations similar to the trend of layering and the eastern contact of the formation in this area.

These breccias are regarded as flow-top breccias, crumble breccias and talus aprons on the flanks of the dome, and the conglomerate as a local epiclastic debris derived from them.

Pyroclastic rocks. Pyroclastic rocks on the northern flanks of the middle rhyolite unit vary from very coarse breccias near the top to tuffs near the northern distal parts. Here, a coarse breccia comprises angular blocks, as much as 2 m long and 50 cm wide, of white rhyolite and grey-weathering rhyolite (or dacite) in a matrix of angular lapilli stone and quartz crystal-bearing tuff. The breccia unit grades over 7 m into a finer lapilli tuff containing almost entirely fragments of rhyolite. Breccias and lapilli tuff are poorly sorted massive rocks in which clasts have a crude preferred orientation (Fig. 33). In one place crude bedding is suggested by 1) change in weathering colour from buff to grey, 2) by preferred orientation of elongate blocks and 3) by distribution of large rhyolite blocks along a particular horizon.

Most pyroclastic rocks are nonwelded, recrystallized vitric-crystal tuffs and lapilli tuffs. The tuffs, although variable in proportions of their constituents, comprise 15 to 20 % whole crystals, and fragments of crystals of quartz, plagioclase (albite), microcline, clasts of recrystallized rhyolite, and relict shards. Crystal material (up to 3 mm across) is almost entirely broken crystals and angular fragments of crystals. Euhedral or subhedral crystals are rare. Shards have blocky, angular, platy and triangular forms that have some concave sides. Vesicular shards and delicate wispy shards are absent. Some rocks contain relict spherulitic texture as seen in thin section. Relict spherulites have

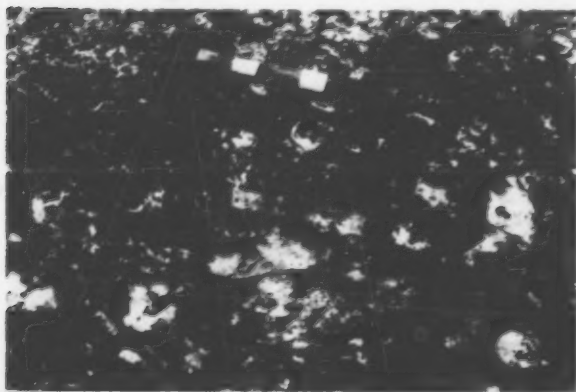


Figure 32. Rhyolite boulder conglomerate in rhyolite complex north of Sunset Lake. Hammer is 33 cm long. GSC 170700

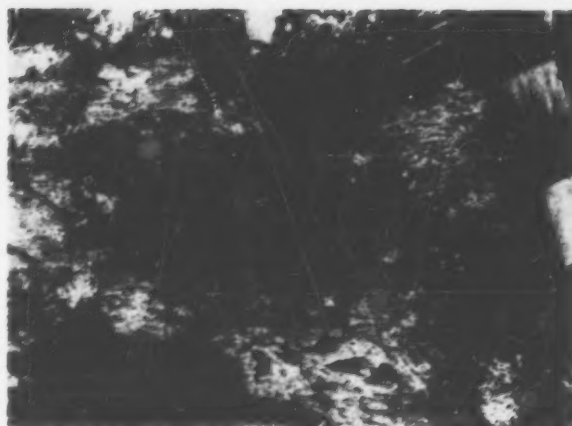


Figure 33. Lapilli tuff in rhyolite at north end of Sunset Lake subarea. GSC 170703

a snowflake texture (Anderson, 1969; Lambert, 1974) in which patches with serrated margins form 3 to 5 sectors within the spherulite. Preserved within the patches is a radiating fibrous structure defined by slender zones of contrasting relief. In some tuffs, shards and pumiceous material contain relict spherulitic texture (snowflake texture) in contrast to areas of crystallographic felsite between shards. Tuffs contain variable degrees of alteration to sericite and calcite. Intensely sericitized rocks are yellowish-grey to greenish-grey with a resinous lustre.

The western unit of rhyolite is a poorly sorted clastic rock comprising subrounded to rounded fragments of rhyolite (up to 3 cm across) and a relatively minor amount of angular fragments of quartz and feldspar. The matrix is very fine granoblastic microfelsite that has an even texture except for abundant alteration to sericite and carbonate. No distinct pyroclastic textures were recognized in thin section, although outcrops resemble tuffs. Dominance of subrounded rock fragments amongst the clasts favours the interpretation that these rocks are epiclastic volcanic sediments.

Dykes and sills. Numerous dykes and sills of rhyolite cut andesitic rocks in an area about 3 km northwest of the north end of Sunset Lake. These intrusions are usually fresh-looking porphyritic rhyolites and dacites. Typically they contain 20 to 30 % phenocrysts of quartz and plagioclase (up to 5 mm across) that occur as single crystals or glomeroporphyritic clusters in an aphanitic matrix. Unbroken, whole phenocrysts are euhedral, subhedral and rounded and embayed, and have very sharp boundaries that are not intergrown with the matrix. Plagioclase (oligoclase to andesine) crystals are unzoned except on very thin margins. The matrix is an even grained, microfelsic mosaic of quartz and feldspar that contains accessory biotite, chlorite and rarely, allanite. In contrast to tuffs and flow rocks these dykes are completely structureless and contain no clastic materials or flow layering. A similarity of phenocryst types in the dykes and in the tuffs suggests consanguinity.

Interpretation

Thick local accumulations of rhyolite represent eruptive centres from which the last gasp of the volcanism took place in the Sunset Lake subarea. Domes developed over a period of time during which volcanic products of intermediate and silicic compositions overlapped and interfingered.

Compositional zonation of the dome south of Sunset Lake indicates rapidly changing composition of the magma. The lower portion of the dome (some of which may be pyroclastic) may be related to felsic andesites and dacites that both underlie and overlie the dome on the northern side.

The dome complex north of Sunset Lake probably had a complicated history of intrusion and extrusion. Distribution of pyroclastic deposits and massive domes and flows suggests that the early stages of volcanism was dominated by explosive activity, whereas quiet effusion of lava or the rise and expansion of domes dominated later stages. The pyroclastic rocks were deposited a considerable distance from these sources, whereas lavas formed thick local accumulations.

Thick massive units, like that near Lake 1052, are probably much more complex than portrayed here (they have not been mapped in sufficient detail to determine their internal complexity). The similarity of stratigraphic setting of this body with rhyolite near Rex Lake suggests that the two bodies may be part of the same unit.

Subaqueous environments are indicated by pillow lavas of intermediate composition that interfinger with parts of the rhyolite members. However, the proximity of the domes south of Sunset Lake with welded tuffs (Alice dacites) suggests that the top of the dome may have emerged above sea level. A subaerial or shallow water environment is also indicated by rhyolite conglomerate and dominance of tuffs amongst flanking rhyolite deposits and overlying felsic andesites. Emergence, if it occurred, was temporary for the rhyolite members were covered with apparent conformity with greywacke and siltstone successions of the Burwash Formation.

Payne Lake Formation

The Payne Lake Formation forms most of the southeastern arm of the volcanic belt in the Sunset Lake subarea (Fig. 21). The formation comprises a complex succession of volcanoclastic sediments that are dominantly layered schists and gneisses in contrast to the Sunset Lake Basalt to the north, which is dominantly pillow lavas, massive lavas and associated breccias. The succession is divided into 3 main lithologies: (1) amphibole-rich schists with variable amounts of mica, that are sediments derived from mafic volcanics, (specific gravity in the range of 2.79-3.00), (2) biotite-muscovite-quartz-feldspar schists that are sediments derived from felsic volcanics; (3) quartz-feldspar schists with minor mica that are regarded as metarhyolite, or detritus derived from rhyolite (specific gravity 2.59-2.68). Intravolcanic sediments are similar to sediments of the Burwash Formation (specific gravity 2.67-2.76).

Amphibolite dykes are absent from the Payne Lake Formation, in contrast to the Sunset Lake Basalt where they form dense swarms intimately associated with the volcanics.

Although the internal structure of this part of the volcanic belt has not been deciphered in detail, the Payne Lake Formation appears to occupy a southerly plunging fold belt that is overturned to the west. It is not certain to what extent the multiplicity of units represents interfingering relationships or results from repetition by folding. Because of the incomplete structural data the thickness of this formation has not been determined and the fold interpretations in Figure 3 are speculative.

Contact relations

Southeast of cross-section P1-P2 (Fig. 3) contacts between the Payne Lake Formation and the Burwash Formation are conformable. Volcanic sediments at locality P1 (Fig. 21) contain coarse conglomerates at this contact. The Burwash sediments tend to have lower relief compared to the volcanic sediments which form a line of bluffs and hills at the contact between the two formations.

North of cross-section P1-P2 (Fig. 3) the Payne Lake sediments lie above the Sunset Lake Basalt. Along the western side of this arm of the volcanic belt the Burwash Formation passes continuously from the Payne Lake Formation to the Sunset Lake Basalt. In doing so it transcends stratigraphic levels.

An intrusive relationship between granitic rocks and the volcanic belt south of Payne Lake is suggested by pods and screens of metavolcanic rock within the cataclastic granite, and by granitic dykes that have intruded roughly parallel to foliation in volcanics.

South of Payne Lake a zone of mylonite 200 m wide and at least 1000 m long occurs between metasediments of the volcanic belt and the granitic terrane (locality P2, Fig. 21). Layering due to shearing, ranging from 1 mm to 15 cm thick, is defined by alternating pale grey to dark grey bands that have sharp contacts. Dark layers vary from even, planar structures to thin, tapering lenses. Dark layers are the zones of deformation and comprise a myriad of microscopic lenticles separated by fine mortar of crushed felsite and oriented muscovite. Felsitic layers are granoblastic aggregate (crystals 0.1-0.3 mm) of quartz, plagioclase, microcline, and minor biotite and epidote. This rock differs from metarhyolitic rocks in other parts of the volcanic belt in that the grain size is coarser and feldspars (of the fine matrix aggregate) are well twinned (they are untwinned or rarely twinned in rhyolite). The mylonite, however, does not contain relics or augen of coarse grains to suggest that it was derived from a granitic rock. The original rock could have been a mylonitized aplite or a coarsely recrystallized felsic volcanic rock (or dyke) of the Payne Lake Formation.

Mafic volcanoclastics, which make up most of the formation, are conformable with other members of the Payne Lake Formation. Near the northern end of the formation this member grades into coarse breccia and pillow breccia of the Sunset Lake Basalt. In these areas, the boundary between Sunset Lake Basalt and the mafic volcanoclastic rocks of the

Payne Lake Formation is drawn arbitrarily where pillow breccia passes into dominantly layered rocks. Such boundaries are approximate. Because of large areas of sparse outcrops and lack of detailed mapping in this part of the volcanic belt, the extent of some members is highly interpretive.

Stratigraphy and lithology

Mafic volcanic sediments. Mafic volcanoclastic metasediments make up 70% of the Payne Lake Formation. The member comprises bedded amphibolite schists that generally do not contain pillow lavas and pillow breccias. No massive lavas have been identified. The mafic composition is reflected by high specific gravities in the range of 2.78-3.06.

In most places this member is well layered. Layers range from thin laminations (0.2-1.0 cm) to beds 80 cm thick. Thin- to medium-bedded rocks are most common. Laminated units are generally even and continuous. Layers of different mineral compositions, contrasting weathering colour, or differential weathering, define thin beds. Discontinuous lenticular beds are not common. Thin beds have sharply defined boundaries, whereas thick beds tend to be massive, crudely layered, and have indistinct boundaries. Attitude of beds generally is steep (65-90°) except near the southern end of the formation where dips are as low as 40-50°. Schistosity is parallel to bedding where bedding is steep, but is steeper where bedding is shallow at the south end of the belt.

Most of these rocks are dark grey to dark green amphibole-rich schists. They contrast with sediments of the Burwash Formation by having abundant amphibole, much less mica and high specific gravity. Hornblende makes up as much as 20 to 45% of the rocks and biotite is generally less than 5%, but in places it is the dominant mafic mineral. Locally these rocks may contain up to 15% epidote. Opaque minerals (magnetite and hematite) are much more abundant (3 to 4%) in mafic volcanic sediments than in other members of the formation, where they make up from trace amounts to 2%.

These rocks are almost completely recrystallized and typically have the metamorphic assemblages:

hornblende-biotite-quartz-feldspar
epidote(sphene)-biotite-garnet-quartz-feldspar

Near the locality P3 (Fig. 21), however, the rocks are anomalously pale grey felsites. These rocks are tremolite amphibolites that have mineral assemblages:

tremolite-quartz-feldspar
tremolite-biotite-epidote-garnet-quartz-feldspar
hornblende-tremolite-biotite-quartz-feldspar

Adjacent beds commonly have contrasting or distinctive metamorphic mineral assemblages. Thin beds (0.2-5 cm), about 1 km east of P1 (Fig. 21) for example, are sharply defined by distinct assemblages of mafic minerals in each layer, namely (1) hornblende only, (2) hornblende-tremolite, (3) tremolite-biotite, or (4) garnet-tremolite-biotite. The contrasting metamorphic assemblages are interpreted

as differences in compositional layering of the original beds.

At locality P4 (Fig. 21) a thin-bedded succession comprises alternating felsic and mafic layers varying from even continuous beds to lenticular units that in some places are present as thin felsic wisps among mafic laminae. The felsic beds are meta-quartz-wackes containing poorly sorted, subrounded to angular quartz and rarely, recognizable relict feldspar detritus. The mafic beds are quartz-feldspar amphibolites that contain relict detrital feldspar and quartz. Hornblende in these rocks is entirely of metamorphic origin and never occurs as detrital grains. Crystalloblastic hornblende that poikiloblastically encloses quartz generally has strong preferred orientation.

Detrital quartz and feldspar in these rocks is recognized by angular, subrounded oval and rounded forms that are generally coarser than the surrounding granoblastic felsic aggregate of quartz and feldspar that is of metamorphic origin. Detrital quartz may be optically continuous single grains, but more commonly they have partly polygonized margins, and in some cases are present as round cores surrounded by irregular marginal overgrowths. Detrital plagioclase varies from turbidity altered (saussuritized) grains to angular, blocky and rounded pseudomorphs that are completely replaced by sericite. Metamorphic plagioclase in granoblastic aggregates is generally clear crystalloblastic grains that are less altered than detrital plagioclase, and in some rocks are coarsely recrystallized to an even granoblastic texture. Detrital plagioclase, however, may still be recognized as rounded cores in crystalloblastic plagioclase. The detrital cores are distinguished from marginal overgrowths by (1) slight difference in extinction angle, (2) quartz inclusions that are present in margins but not in cores, (3) trains of minute dust-like inclusions that mark the boundary between core and margin, and (4) change in development of twinning. Overgrowths have well developed polysynthetic twinning, whereas cores are untwinned or twin units that taper out where they pass from the marginal overgrowth to the cores.

Clastic textures are not obvious megascopically except where fragments are coarse. Coarse fragments are commonly deformed.

Some thin-bedded mafic rocks that lack any vestige of clastic texture are very fine even granoblastic aggregates of quartz and feldspar, with lepidoblastic biotite and poikiloblastic garnet and hornblende. Such rocks may be metavolcanic lutites derived from a mafic source.

Metarhyolite. Rocks of rhyolitic composition form a northerly trending band, about 8 km long, down the north-central part of the formation and pairs of bands of unknown extent, but about 200 m wide, near the eastern side of the formation. Similarity in structure and lithology of pairs of rhyolite units along the eastern side of the belt, as well as pairing of the sedimentary units, suggests that repetition of units resulted from folding. The rocks range from massive rhyolite to coarse rhyolite breccias in the northern part of the central band, to deformed breccias and layered arenites in the middle and southern parts.

Breccias are invariably deformed coarse lenticular rocks and comprise strongly oriented white weathering felsic clasts in a dark grey micaceous, schistose matrix. In some cases the matrix is tightly packed, fine lenticles separated by fine grained biotite and muscovite. Coarse clasts make up 30 to 75 % of some outcrops. They vary from smooth, tapering lenses and gently undulating streaks, to elongate blocks with rounded to angular, blunt, stubby ends, and range from 1 to 5 cm wide and 15 to 50 cm long. Thin wispy streaks typify outcrops where the amount of clasts is low, whereas stubby and blocky forms are common where concentrations are high. Rocks with abundant thin tapering lenses resemble eutaxitic foliation in welded tuffs.

The middle and southern parts of the central band contain massive to well layered felsic meta-quartz-feldspar arenites. Although the rocks are recrystallized, rounded to ovoid grains of detrital quartz are preserved. Some grains have irregular margins that are thin overgrowths on ovoid cores. Schistosity and orientation of clasts are generally parallel to layering in finer clastic units. Layering in felsic volcanics is folded and partially sheared.

Massive rocks and coarse breccias are interpreted as lavas or sills. Breccia is mainly coarse debris flanking the lavas, but some may be tectonically deformed sills or dykes. Fine layered rocks and arenites represent accumulations of fine epiclastic detritus.

Felsic volcanoclastics. Felsic volcanic sediments form several tongues or lenses within the northern half of the Payne Lake Formation. These buff to medium grey felsites are phyllitic, or weakly schistose, and contain a lesser amount of biotite (generally less than 20 %) and tend to be more siliceous than the metavolcanic sediments or the sediments of the Burwash Formation whose biotite content ranges up to 40 %.

Most outcrops show evidence of clastic texture. Bedding varies from poorly defined, thick massive layers, to clearly defined, thin-bedded successions.

The northernmost unit varies from massive, coarse breccia near the north end to massive and well bedded volcanic arenites in the southern parts. Some of the coarser breccias and volcanic arenites contain more mafic minerals than the normal felsic volcanic sediments, and may be fine sediments derived from adjacent mafic pillow breccia sequences. The middle and southern parts of this unit contain coarse arenites and fine rudites in which the felsic clasts are deformed into lenses (commonly 2-5 mm wide and 5-10 mm long, and locally up to 5-10 cm long) that are oriented parallel to schistosity in the matrix. The very fine grained, sugary textured, pale grey matrix is completely recrystallized to an aggregate of quartz, plagioclase and mica. Fine laminations (1-5 mm thick) in layered units are defined by contrasting grain size, variable amounts of mica and differential weathering. One schistose unit in this member has the assemblage: biotite-hornblende-staurolite-cordierite-quartz-feldspar. The appearance of cordierite and staurolite is anomalous in this area in both sediments and volcanic rocks.

The eastern bands of felsic volcanites comprise massive to thin layered units. Layers averaging 0.5-4 cm thick and locally up to 25 cm thick commonly have diffuse boundaries. Some units near the northern end contain abundant actinolitic-hornblende amongst granoblastic aggregates of quartz and feldspar. In spite of the high amount of mafic constituent (about 35-40 %) specific gravity of this rock is relatively low (about 2.75).

Rocks in the northern parts are muscovite-biotite schists and have minor hornblende. The areas between lepidoblastic micas are granoblastic aggregates of quartz and plagioclase (oligoclase) and minor garnet and tourmaline (schorlite), calcite and chlorite. Relics of quartz grains 0.5-1 mm across are oval to round, coarse-polygonized aggregates that have mortar texture that tapers off in the plane of schistosity.

This unit is interpreted as detritus derived from felsic volcanic units and local mixing of detritus from mafic volcanic sources. Some rhyolite volcanoclastics are almost identical to the adjacent massive rhyolite units to the east from which they were probably derived.

Intravolcanic sediments. Areas of buff, grey to pale brown metasediments form large units near the southeastern and western sides of the formation and numerous small units generally in the southern part of the belt. These rocks are mica schists that lack hornblende, like sediments of the Burwash Formation. Their aluminous nature is indicated by the presence of garnet, staurolite (rare) and abundance of mica. Metamorphic mineral assemblages include:

biotite-quartz-feldspar
garnet-biotite-quartz-feldspar
biotite-muscovite-garnet-staurolite-quartz-feldspar

Crystalloblastic plagioclase is oligoclase in some rocks. One staurolite-bearing rock contains reversely zoned andesine that ranges from An_{37} to An_{56} (measured in different crystals): anorthite content of many crystals differs by about An_5 from core to rim. These rocks tend to be more coarsely recrystallized than the volcanic sediments of other members, hence, relics of detrital quartz and plagioclase, although present, are not easily distinguishable. These units may represent pockets of pelitic sediments within the volcanic belt, or part of the Burwash Formation that has been infolded into the belt.

Interpretation

The Payne Lake Formation is interpreted as volcanoclastic sediments derived directly from the adjacent volcanic pile. Since these are all immature sediments, specific gravities are interpreted to reflect compositions of a parent rock from which they were derived. The mafic member is considered an epiclastic facies of the Sunset Lake Basalt into which they grade. Alternating mafic and felsic beds may represent provenance from a mixed source or variable conditions during sedimentation, such that unstable mafic constituents were removed from the detritus. Some of the rhyolitic units may be subaqueous flows or domes from which coarse rhyolitic blocks formed proximally and fine clastic sediments

formed distally. The felsic member is assumed to represent detritus derived from a felsic volcanic source and partly from rhyolitic units. Intravolcanic sediments could represent pelitic pockets within the volcanic belt, fingers of the Burwash greywackes near the distal margins of the volcanic sediments, or synclinal keels of the overlying Burwash Formation. The environment of deposition is considered to be entirely subaqueous.

Carbonates

Carbonate units locally mark the contact between felsic volcanics and sediments of the Burwash Formation in all three subareas. Main units occur between Turnback and Rex lakes, near Lake 1052, and south and west of Detour Lake. The largest units are in the order of 85 m thick. Carbonate-bearing rocks can be traced intermittently through islands in Detour Lake and along thin felsic volcanic units between Detour and Upper Ross lakes. Carbonate occurs as lentils within rhyolite units near Lake 1052 and at the contact between rhyolite and the Alice andesites between Sunset Lake and Lake 1052. It forms the matrix of rhyolite breccias and conglomerates associated with rhyolite domes and flows north of Sunset Lake and of rhyolite boulder conglomerates and breccia along the volcanic-sedimentary contact north-east of Ross Lake.

West of Lake 1052 carbonate forms a large lenticular unit 200 m thick and about 1600 m long between felsic volcanics and layered mafic volcanics. It also forms a 30-50 m thick zone, within both mafic and felsic volcanics, that follows the southern side of the volcanic belt at the contact with Burwash sediments. Where carbonate forms the matrix of breccias and conglomerates, it is interspersed within the rocks and does not form discrete units.

Contacts between carbonate units and adjacent volcanics are gradational over distances of 10 to 50 m. In contact zones rocks vary from volcanic breccias containing abundant carbonate matrix to volcanic units having little or no carbonate. West of Lake 1052 the largest carbonate unit tapers out at the north end to a mafic breccia in which carbonate forms the matrix between closely packed mafic lentils over a distance of about 15 m.

Internal structures

Foliation, defined by trains of small fragments, blocks, lenses and slivers of volcanic material that have strong preferred orientation, generally dips steeply. The foliation has the same trend as boundaries of the unit and has approximately the same attitude as layering and foliation in adjacent volcanic units. Smaller fragments commonly have flat, irregular and thin (some only 1 mm thick) discoidal shapes with strong preferred orientation (Fig. 34a) that, where closely packed and aligned, resemble bedding.

Trains of fragments locally outline convoluted layers and minor folds that commonly have the same attitudes and sense of movement as structures in adjacent rhyolite units. In one place carbonate contains boudins of felsic volcanics and pegmatite (Fig. 34c).

Lithology

The large carbonate unit forms ridges above low-lying sediment between Turnback and Rex lakes. Units are almost all breccias or coarse gritty rocks comprising felsic volcanic

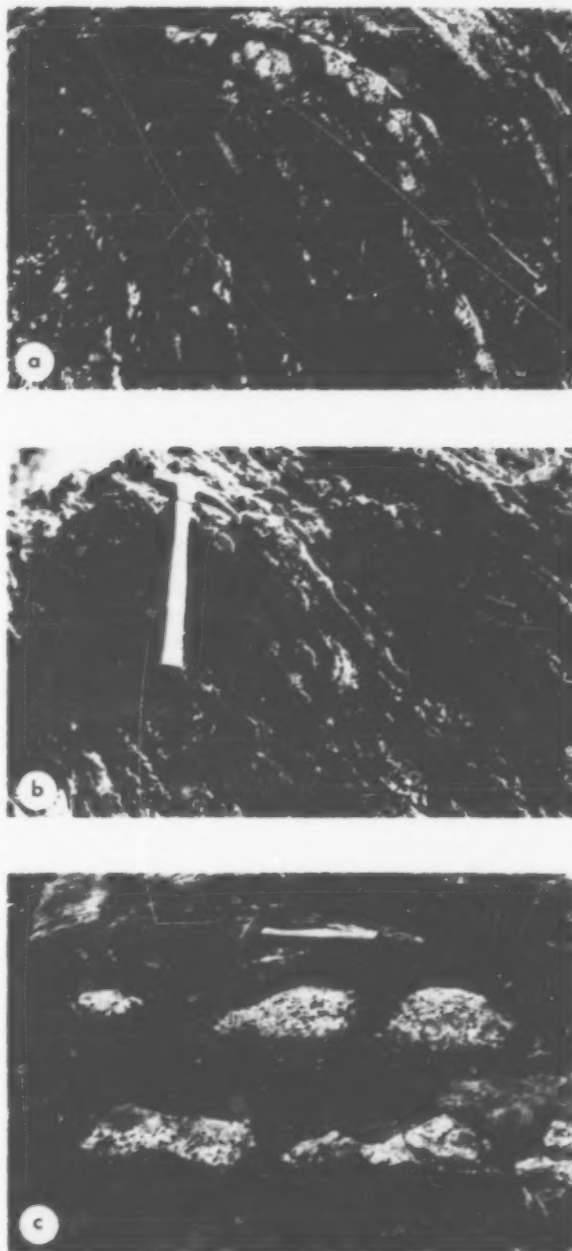


Figure 34. Textures and structures in carbonate unit, north end of Turnback Lake. (a) Strong preferred orientation of lenses and discoidal clasts of rhyolite in carbonate matrix (GSC 170692). (b) Folded rhyolite clast (centre) (GSC 170689). (c) Boudinaged pegmatite in rhyolite-carbonate grit. Hammer is 33 cm long. (GSC 170693).

fragments in a matrix of carbonate grit. Fragments of rhyolite range from 1 mm up to huge screens 600 m long and 50 m wide. Screens of rhyolite several hundred metres long near the south end of the unit, near Turnback Lake, are represented diagrammatically as a single lens in Figure 3. Between Turnback and Rex lakes fragments generally are more abundant near the southern end of the belt where they make up as much as 40% of the rock. Locally, near the northern end, the unit is dominantly black-weathering carbonate and contains only sand-sized grains of crystals and rhyolitic material.

Volcanic fragments within the carbonate are almost invariably flattened to elongate discs and slivers, some of which are fragments of sheared rhyolite between which carbonate has permeated. Where units comprise sand-sized grit, the tiny fragments are thin, irregular discs. Fragments are invariably poorly sorted and oriented parallel to tectonic foliation.

Carbonates weather dark brown to black and fresh rocks are pale to dark grey to dark green, depending on the composition, size and abundance of volcanic fragments. Most commonly, pale grey weathering felsic fragments stand out in relief 1 to 20 mm above the dark brown to black, fine- to medium-grained crystalline carbonate matrix.

Carbonates north of Turnback Lake have a fine crystalloblastic matrix containing dominantly calcite 0.2-1 mm across and lesser amounts of rounded, oval to regular quartz, coarse crystalloblastic clinopyroxene, epidote, actinolite, microcline, sphene, and magnetite, in diminishing order of abundance.

Carbonate matrix of rhyolite conglomerate north of Sunset Lake is a microbreccia comprising angular fragments of porphyritic rhyolite, feldspar and quartz crystals in a matrix varying from extremely fine (0.02 mm) to medium-granular (0.04 mm) dolomite. Veins of calcite cut the rock. Rock fragments and crystals are lithologically identical to the phenocrysts and rhyolite of the body with which the conglomerate is intimately associated. The carbonate contains fine veins of granoblastic quartz and coarse (2 mm grains) carbonate. Some rhyolite fragments are almost completely replaced by very fine granular carbonate. This rock does not contain the pyroxene, amphibole and epidote assemblages seen in the unit north of Turnback Lake.

Carbonate on islands in Detour lake and between Detour and Victory lakes are carbonate-rich zones in rhyolite breccias, amphibolite schists and schistose metasediments. The carbonate-rich units are cut by numerous small granitic intrusions and granitic dykes that have well developed zones of skarn at their contacts. Numerous small plutons that intrude the carbonate unit north of Turnback Lake have haloes of coarse diopside, garnet and skarn up to 7 m across. These skarns contain poikiloblastic diopside and garnet, forming crystals and clusters of crystals 1 to 5 cm across, quartz and minor carbonate.

Pyrite-rich gossans form along contacts between carbonate and the Turnback Rhyolite and the Burwash Formation.

Interpretation

Carbonate units do not form discrete beds (sedimentary layers) in the volcanic-sedimentary succession of the Cameron-Beaulieu volcanic belts. Instead, they are carbonate-rich zones in clastic volcanic rocks (that may be primary breccias or tectonic breccias) mainly at the top of rhyolite units where they are overlain by the Burwash Formation. Carbonate units are not found in the Burwash sediments overlying the volcanic succession. The carbonate was emplaced during the late stages of volcanism or after volcanism.

That the carbonate was deposited in coarse clastic volcanic rocks by volcanic emanations (exhalative deposits) during the waning stages of volcanism, is consistent with the following features of the carbonate units: (1) presence of carbonate, mainly in breccias at the top of late-stage rhyolite units, some of which are domes; (2) gradation of contacts with adjacent volcanic units; (3) a general parallelism of structures in carbonates with those in adjacent volcanics, and (4) sulphide zones concentrated on the upper sides of the carbonate unit that makes contact with the Burwash sediments. During folding and tectonic deformation, carbonate (being less competent than the felsic rock in which it permeated) deformed by plastic flowage, whereas the volcanic material behaved in a brittle fashion to become further fragmented, boudinaged and drawn out into trains of fragments.

Dykes and sills

Mafic intrusions form: (1) amphibolite dykes and sills within the volcanic belts as single intrusions or dense swarms; (2) amphibolite dykes swarms in the western margin of the Sleepy Dragon Complex, adjacent and parallel to the Cameron River belt, and in granitoid rocks along the north-eastern side of the Beaulieu River belt; (3) highly deformed amphibolite layers of variable attitude within the Sleepy Dragon Complex; and (4) Proterozoic diabase and gabbro dykes that cut across basement, supracrustal and plutonic rocks.

Felsic intrusions form rare dykes in the volcanic belts associated with rhyolite-dacite dome complexes and late northwest-trending felsite intrusions that cut across both volcanic and sedimentary rocks of the Yellowknife Supergroup.

Late cross cutting felsite dykes and diabases are probably all Proterozoic and unrelated to the present volcanic belts. Likewise the amphibolite layers and lenses that form tight intrafolial folds within granitoid gneisses of the Sleepy Dragon Complex (Davidson, 1972) probably are older than the volcanic belts.

'Amphibolite intrusions' are diabase to gabbroic dykes and sills metamorphosed to amphibolite grade. Amphibolite 'dyke' is used here as a loose term for all premetamorphic mafic dykes and sills.

Amphibolite intrusions within parts of the volcanic belts are considered to be intimately related to development of the volcanic piles.

The dyke swarms in the Sleepy Dragon Complex, however, may not be directly related to the adjacent volcanic belts nor the dykes they contain as previously interpreted (Baragar, 1966; Lambert 1977, 1982). Amphibolite dykes have not been traced across the complex zones of high strain that separate the Sleepy Dragon Complex from the volcanic belt in the Cameron River subarea nor the dyke-granitoid complex of the Beaulieu River belts along the northeastern side of the Sunset Lake subarea (Fig. 35). Dykes in the Sleepy Dragon Complex do have some lithological and structural differences from those in the volcanic belt.

Amphibolite intrusions within volcanic belts

Distribution and thickness

Amphibolite intrusions within the volcanic belts variably form (1) dense swarms in and parallel to the Cameron River belt (Fig. 35a), (2) large complex dykes or sills within the northwestern and northeastern margins of the volcanic belt in the Sunset Lake subarea, and (3) sills in various parts of the Tumpline Lake subarea. With one exception southwest of Tumpline Lake, they do not intrude the Burwash Formation. Mapping of these swarms is incomplete (shown by the variable detail in Figure 2: no detailed mapping was done in the northern third of the Cameron River belt) and of insufficient detail, except in local selected areas, to subdivide the swarms. Consequently, the following account gives a rather generalized description of the intrusions.

The greatest concentration of dykes generally is along or near margins of the volcanic belts where they make contact with granitic terrane. The intrusions are almost parallel to the general trend of the volcanic belts, internal stratigraphy and trend of flattening of pillow lavas. In places, dykes form higher and rougher topography than the surrounding rocks. West and southwest of Amacher Lake, for example, amphibolite stands as a prominent ridge 70 m above the generally lower lying granitic terrane to the west.

In the Sunset Lake subarea amphibolite intrusions are restricted mainly to the Sunset Lake Basalt and tend to become thinner and sparser towards the top of the formation. They do not penetrate the Alice Formation, Payne Lake Formation or rhyolite. West and south of Amacher Lake amphibolite dykes generally less than 400 m wide, trend north-northwesterly parallel to the trend of lithological units and the boundary with the granitoid gneisses to the west. They tend to form large complex intrusions rather than dense swarms like those in the Cameron River belt.

In the Tumpline Lake subarea amphibolite dykes and sills form large sheets near Tumpline Lake, south of Detour Lake and west of Sharrie Lake. The largest intrusion occurs along the contact between Tumpline Basalt and the Turnback Rhyolite near the northwestern side of Tumpline Lake. This body, shown as three units in Figure 2, is a single sheet that has been folded. The westernmost exposure occupies the crest of a doubly plunging, isoclinal fold that is overturned to the northwest. The intrusion is at least 200 m thick (possibly as much as 400 m) and is exposed for 8 km.

Amphibolite dykes and sills are notably sparse in the easterly trending wings of the volcanic belt (northeast of Tumpline Lake) where most of the rocks are dominantly volcanic sediments.

Near Sharrie Lake dykes intrude both basalts and rhyolites. Locally there is no clear distinction between sills and possible lavas that follow the contact between the two units.

In the Cameron River subarea swarms of amphibolite dykes and sills generally make up 20 to 35 % of the volcanic belt (Fig. 2, 8, and 35). Thickness of dykes varies from a few tens of centimetres to 200 m: most are between 20 and 100 m. Some large intrusions in the volcanic belt, that have been traced for 6 km, pinch and swell and occasionally split into two or more thinner units that surround lenses of pillow lava up to 100 m thick.

Contact relations

In the Cameron River belt large dykes and sills have intruded along contacts between units and generally are almost parallel to the direction of pillow flattening and to the trend of the volcanic belt. The intrusive nature of these bodies is indicated by:

1. Contacts that cut across pillows (a feature not commonly well documented because of the high degree of deformation).
2. Sharp, regular contacts that lack surface structures typical of lava flows.
3. Relict contact metamorphic aureoles defined by coarsening of grain size of amphibolite pillow lavas adjacent to contacts with dykes.
4. Relict chilled margins indicated by zonal textural variations ranging from aphanitic through fine- to coarse-grained, gabbroic-textured centres.
5. Inclusions of fine grained amphibolite in margins.

Generally large dykes are complex intrusions. Internal chill contacts suggest several generations of intrusions where late dykes split off parts of earlier dykes so that one-way chill margins on dyke fragments are common.

The large intrusion along the western side of the Cameron River belt shows clear intrusive relations with pillow lavas where it is thick. Locally near the northeastern end, where it thins, the contact between massive amphibolite and pillows is gradational and the massive dyke "buds" into pillows having the same texture as the dyke. These features document the transition where dykes have surfaced and effused as lavas or, in some cases, where massive sheet flows change to pillowed flows. Such transitions are present only locally along the length of some dykes.

Pillows at the contact of a large sill west of Webb Lake are flattened at the immediate contact of the intrusion, whereas farther away they are equant in horizontal section. Although such flattening may have taken place during emplacement of these synvolcanic sills (possibly where pillows were not completely solidified) it is equally possible that the change in style of deformation is related to the local variation in the pattern of strain where relatively incompetent pillow lavas deformed against competent and rigid intrusions.

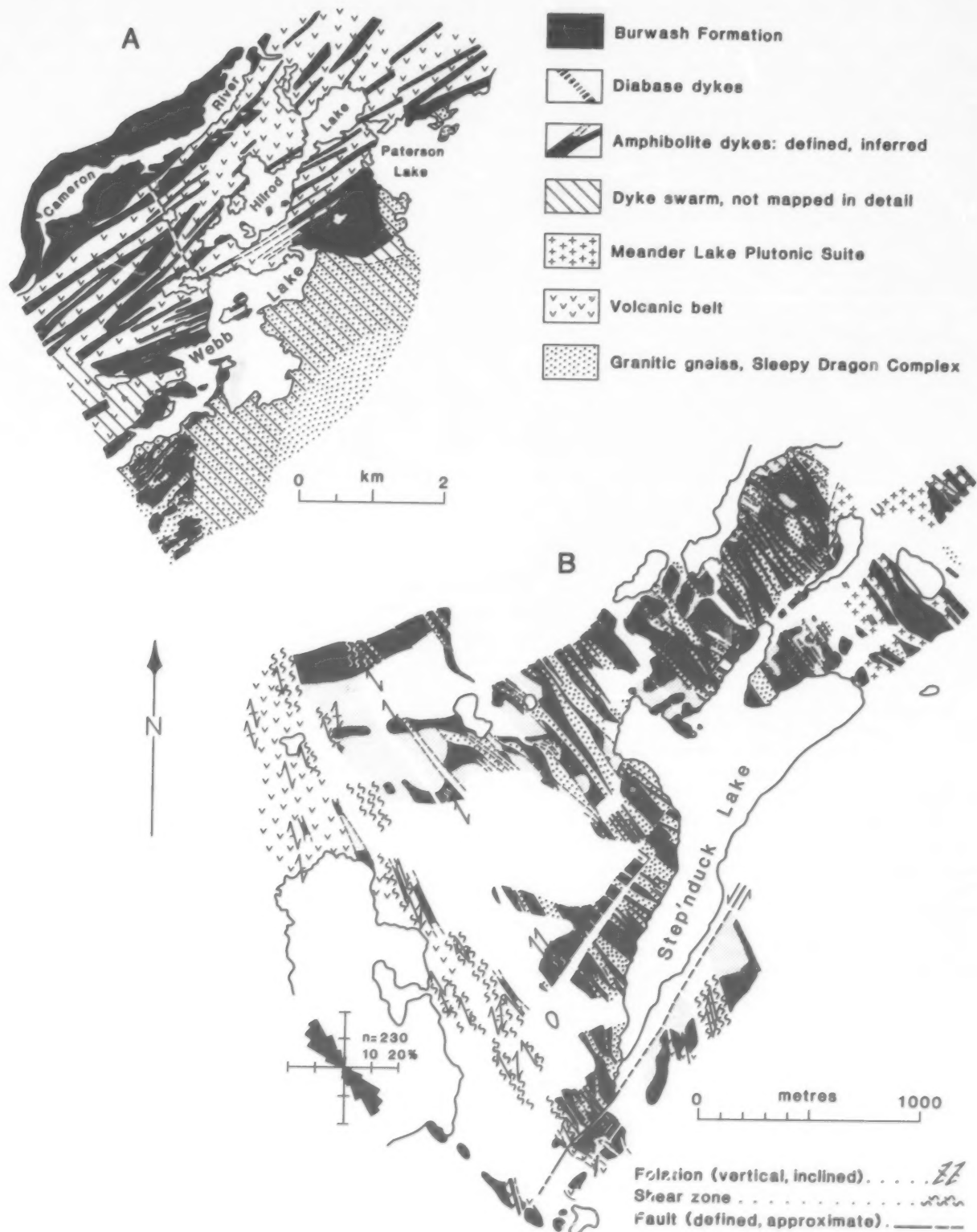


Figure 35. Amphibolite dyke swarms in (a) Cameron River and (b) Sunset Lake subareas. Map locations are shown in Figure 51.

Where deformation in pillow lavas is mild sharp cross-cutting and minutely irregular dyke contacts are identifiable. For example, at one locality within the easterly trending wings of the volcanic belt (northeast of Tumpline Lake) sills of amphibolite have small dykelets projecting into layered volcanic rocks on either side. Contact relations are difficult to establish in areas of intense deformation.

South of Detour Lake large amphibolite sills are surrounded by zones, 50-100 m wide, of mixed amphibolite and felsic volcanic material. Margins include equant to elongate blocks of dominantly felsic volcanics, but also some granitic material, ranging from a few centimetres to 1.5 m across. In one place a screen of rhyolite is 10 m wide. In places the inclusion-rich margins of dykes grade outward into a mixed zone. The mixed zone varies from swarms of small amphibolite dykes, gneissic rocks consisting of conformably interlayered felsic volcanics and amphibolite (0.3-1.0 m thick), to a myriad of fine layers and irregular lenses of amphibolite and felsic volcanics. In some cases these areas of mixed rock are interpreted as zones of magma injection around large dykes that intruded a highly fractured area in the rhyolite. Subsequent deformation may have caused attenuation and separation of small bodies into pods and lentils. In other places the mixed zone could be highly strained inclusion-rich margins or margins that have suffered tectonic disruption.

West of Amacher Lake north-northeast-trending mafic intrusives parallel the complex boundary zone between the Beaulieu River volcanic belt and the Sleepy Dragon Complex to the west. Although different generations of intrusions have not been distinguished in this area, early deformed bodies and late dykes that cut across major structures have been recognized. A 6 km long by 300 m wide amphibolite dyke intrudes or incorporates all units of the boundary zone. Although discordant relationships are locally preserved, the thin large intrusive complex may not have been a true dyke but rather a sill. This is a complex multiple intrusion, made up of several units as indicated by internal zones of fining and chilling, that, in its northern part, completely encloses a unit of iron-formation.

Near Amacher Lake, contacts with mafic volcanics locally are characterized by well developed foliations and/or lineations and commonly are the locus of gossans. Contact with granitoid gneiss bodies are generally obscured by deformation so that chilling relations are difficult to establish.

Some rocks at the immediate contact with the amphibolite intrusions show effects attributed to contact metamorphism. Sediments and felsic volcanics adjacent to the large intrusion along the north side of Tumpline Lake, for example, contain unusually coarse cordierite, garnet and mica. Metarhyolites (or felsic volcanic sediments) adjacent to the western side of the dyke west of Tumpline Lake are coarse grained garnet-biotite-muscovite schists containing garnet up to 2 cm across. Metasediment incorporated into a dyke is cordierite-mica schist with cigar-shaped crystals of cordierite up to 10 cm long and 2 cm wide. Although contact metamorphic effects are rarely recognizable in mafic lavas, grain size of amphibolite pillow lavas becomes coarser at

the immediate contact with some large sills in the Cameron River belt.

South and west of Detour Lake, gossan zones rich in pyrite and pyrrhotite occur within margins of mafic dykes and in the surrounding felsic volcanics. Similarly, west of Amacher Lake, gossans containing pyrite, pyrrhotite, chalcopyrite and magnetite are common along contacts between amphibolite dykes and the Sunset Lake Basalt.

Internal structures

Internal chill contacts (discussed previously) generally parallel the trend of main complex intrusions although small late dykes may wander through larger host dykes. Simple dykes are massive throughout, except for subtle layering defined by gradual fining of grain size towards contacts. Some coarse grained dykes have pronounced, steeply plunging mineral streaking defined by preferred orientation of coarse hornblende and in places by felsic constituents. Locally mineral streaking parallels dyke margins.

Layered margins are common in thick intrusions. The huge folded dyke along the west side of Tumpline Lake, for example, is well layered from 10-30 m from the contact (in the westernmost exposure) and in one place possibly as much as 100 m from the contact. The eastern exposures of this dyke, however, are generally massive with only minor layering at the immediate contacts. The diffuse layering is defined by continuous to streaky segregations of plagioclase ranging from 0.2 to 10 cm wide. In one place the layers are locally deformed into flow folds. Generally, the layered amphibolites are fine- to medium-grained but some coarse grained parts are also layered.

Massive interiors of intrusions commonly have a blocky, square to rectangular pattern of joints. Within the margin of one dyke northwest of Webb Lake joints make an angle of 30-35° with one another to produce a pattern of elongate trapezoids parallel to the dyke margin. In places these trapezoids superficially resemble flattened pillows.

A large dyke west of Amacher Lake superficially appears massive and relatively undeformed, but high strain is indicated by a pervasive foliation, fine grain size (reduced from presumed coarse grained size typical of large dykes elsewhere in this area), folded felsic veins, and intensely deformed migmatitic margins where locally the dyke has included swarms of granitoid inclusions. The folding of the enclosed iron-formation also suggests considerable deformation.

Lithology

Mafic intrusions in all belts form homogeneous massive amphibolites that weather dull brown, rusty brown, grey-green to buff-green and dark grey. Dykes typically have medium-to coarse-grained centres that grade into dark grey, almost black, fine grained to aphanitic margins (relict selvages). Layered margins display alternating pale grey to dark grey bands that have diffused boundaries. Weathered surfaces accentuate gabbroic textures in medium- and coarse-grained rocks and diabase textures in the fine grained

rocks. Highly strained dykes are amphibolite schists and are locally gneisses where they have included granitic or felsic volcanic material.

Lithology of the amphibolites is essentially the same in all belts. Local exceptions are in columns 3 and 5 (Fig. 8b) where both dykes and the pillow lavas they intrude are coarse feldspar-phyric rocks. The lithology of the feldspar-phyric rocks is described with the Cameron River Formation.

Amphibolite intrusions within the volcanic belts are completely recrystallized rocks. Metamorphic assemblages typically are hornblende (actinolite)-plagioclase-quartz-epidote and calcite. Amphibole is poikiloblastic, blue-green hornblende that encloses irregular minute grains of crystalloblastic plagioclase and quartz, and in some cases, calcite and epidote. In some medium grained rocks of slightly lower metamorphic grade, where actinolite rimmed with blue-green hornblende dominates, amphibole appears to have replaced pyroxene, grain for grain, and thus preserves the original gabbroic texture. Amphibole in these rocks has sharp, regular to blocky outlines in contrast to the more common amphibolites where hornblende has extremely irregular, ragged outlines, and grains crosscut one another. Actinolite is not poikiloblastic. Plagioclase in most rocks is recrystallized or polygonized to a fine irregular crystalloblastic aggregate. Relics of primary plagioclase are felsic aggregates whose outlines define elongate to stubby lath forms. In porphyritic rocks, however, plagioclase phenocrysts are not recrystallized or are only partly polygonized, in contrast to the completely recrystallized plagioclase in the matrix. Some phenocrysts of feldspar are saussuritized, but maintain their euhedral to subhedral outlines. In general, rocks that have relict textures of coarse grained gabbro are more coarsely recrystallized than the fine grained amphibolites. In coarsely recrystallized rocks, felsic constituents are fairly regular granoblastic mosaics of quartz and feldspar crystals up to 0.1 mm size, whereas in fine grained and aphanitic amphibolite, felsic constituents form an irregular crystalloblastic mosaic.

Felsic inclusions are common in dykes within the Sunset Lake and Tumpline basalts but are not common in dykes of the Cameron River Basalt, except for those along the eastern margin.

The dyke along the northwest side of Tumpline Lake contains inclusions of rhyolite within 30 m of its contact. In one area rhyolite inclusions form slivers 30 to 60 cm wide and 15 m long and equant blocks up to 1.5 m across. Slivers are oriented parallel to layering and dyke contacts.

In the Sunset Lake belt unusual lensoid structures are created where an amphibolite dyke has intruded and incorporated into its margin a maze of screens and fragments of rhyolite. The resulting mixed rock has been intensely sheared and deformed. In this mixed zone rocks vary from fine grained and massive, to sheared amphibolite, to a streaked rock containing oriented pinching and swelling bodies of felsite and white weathering felsic lenticles varying from undulatory tapering streaks to lenses that have irregular minutely digitated terminations (Fig. 36a). Locally the rock is gneissic where it comprises tightly folded felsite and 2 to 10 cm thick mafic layers (Fig. 36b).

Effect of intrusions on structure and stratigraphy

Dykes and sills in the Cameron River subarea, because of their great thickness and lateral extent, have the following local effects on the stratigraphy, they:

1. Offset thin units so that stratigraphic correlation is not obvious.
2. Produce apparent displacement of thin units that could be misinterpreted as fault displacements.
3. Form apparent multiplicity of units by splitting off parts of units or splitting units laterally and separating the parts by up to 100 m.
4. Create apparent interfingering of units.
5. Spread units within a formation so that they appear much thicker than they really are (the lowest tongue of the Webb Lake Formation, between columns 6 and 7, Fig. 8 that appears to be up to 300 m thick but actually is only about 180 m because it is intruded by 120 m of gabbroic sills).

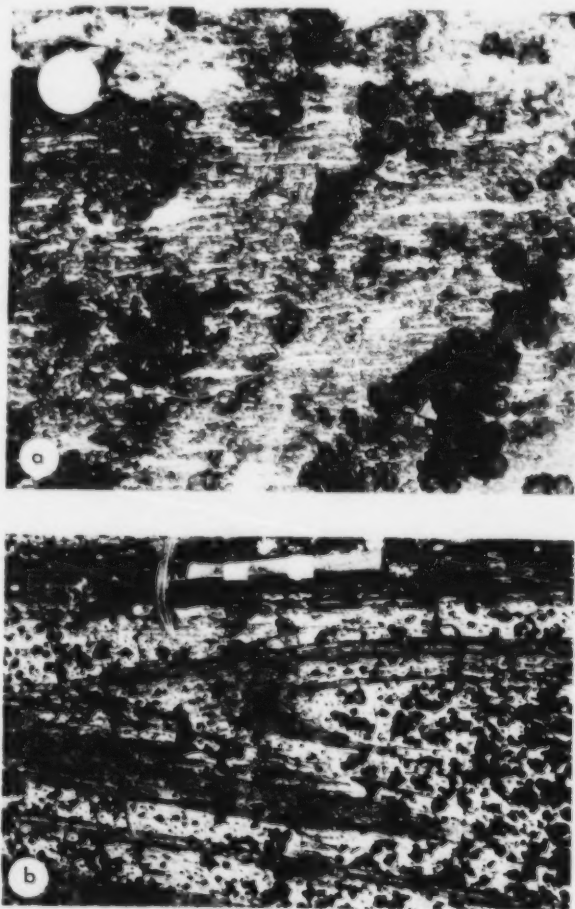


Figure 36. Highly strained amphibolite containing inclusions and screens of felsite 4.5 km east of Sunset Lake. (a) Deformed inclusions produce a pseudo-eutaxitic structure (GSC 170766). (b) Folded felsite-amphibolite "gneiss" with axial planes parallel to trend of major shear zone (GSC 170770).

Removal of sills (graphically) from the volcanic belt generally facilitates stratigraphic correlation in the volcanic succession. However, although removal of sills is necessary for logical correlation within the eastern parts of the belt, the equivalent thickness of these lower sills must be replaced to facilitate correlation within the western part. Hence, at least some sills in the lower part of the belt must have been emplaced before the upper parts of the belt were deposited. The penecontemporaneous emplacement of pillow lavas and intrusions is further suggested by:

1. Areas where sills and pillow lavas have the same distinctive lithologies.
2. Sills or dykes that locally pass into pillowed lavas.
3. The restriction of intrusions to the volcanic belt and their absence in the overlying and adjacent sediments. Thus, the sills are an integral part of the volcanic stratigraphy.

Interpretation

Amphibolite dykes and sills were feeders for the pillow lavas in the various volcanic belts. Some units interpreted as sills may in fact have been sheet flows. During each new eruption, dykes and sills intruded the preceding lavas. Some intruded laterally before emerging as pillow lavas. Probably dykes originally were emplaced as steeply dipping sub-parallel swarms trending in a similar direction to the elongation of the volcanic pile. During subsequent deformation the entire volcanic-intrusive succession was intensely deformed on, or against, the Sleepy Dragon Complex, so that sills, dykes and volcanic stratigraphy tended to rotate into similar trends.

Intrusions in granitoid gneisses

Amphibolite intrusions form impressive, dense dyke swarms in the Sleepy Dragon Complex east of the Cameron River belt and in granitoid gneisses along the northeastern side of the Beaulieu River volcanic belt, and form extensive dykes in the Sleepy Dragon Complex along the western boundary of the volcanic belt in the Sunset Lake subarea. In all places the general trend of dykes in the granitic terrane is essentially parallel to dykes in the adjacent volcanic belts and structurally complex boundary zones between the two terranes.

The impressive swarms in the Cameron River subarea have been outlined by early mappers (Henderson, 1941; Baragar, 1966) and subsequent work (this report; Lambert, 1982; Lambert and Ernst, 1987; Lambert and van Staal, 1987) provided more precise documentation of relationships between swarms in volcanic belts and granitoid basement in both the Cameron and Beaulieu volcanic belts.

Cameron River subarea

In the Cameron River subarea a dense swarm of dykes occurs within the granitic terrane up to 4 km east of the volcanic belt (Henderson, 1941). This swarm has been mapped in detail only in one small area southeast of Webb Lake. Henderson (1985, Fig. 5, p. 9) showed the density of mafic intrusion in the granitoid terrane north of Upper Ross Lake.

Figure 2 shows, diagrammatically, by a hachure pattern, the approximate limit of this swarm.

Dykes in the granitic terrane have less regular trends than in the volcanic belt but still have crude northeasterly orientations. Identification of intrusions is difficult along the contact with granitic terrane because of the ubiquitous extremely high strain which has obliterated primary textures and structures. Hence, individual intrusions have not been traced across the contact between the granitic terrane and the Beaulieu Group.

Dyke intensity profiles across parts of the swarm that have been mapped in detail, and the more generalized description of the swarm by Henderson (1985), suggest that dykes form 10 to 60 % of the dyke-granitoid complex and locally are as profuse as dykes in the Cameron River belt.

Dykes within the granitic terrane east of Webb Lake, in contrast to those in the adjacent volcanic belt, are less intensely recrystallized, fine grained rocks. They make sharp intrusive contacts with the granitic host and commonly have weakly foliated margins. Ophitic textures, in which plagioclase forms euhedral to anhedral laths that are clear, unaltered and not polygonized, are preserved. Cores of plagioclase phenocrysts range up to bytownite (An_{85}) and thin margins have normal or oscillatory zoning. In some rocks, however, plagioclase is almost completely altered to sericite except for thin margins. Mafic minerals are completely recrystallized to blue-green hornblende. Amphibole is not poikiloblastic and in some cases appears to have replaced pyroxene and maintained original sharp outlines. Zoned hornblende has broad cores of pale green, fine, flaky hornblende, whereas rims are deeper blue-green, broad, optically continuous zones.

Dykes invariably contain felsic inclusions. Granitic xenoliths vary from angular equant blocks, irregular slivers and large screens, to rounded, elongate amoeboid and lensoid schlieren. Elongate inclusions have preferred orientation parallel to dyke contacts. Inclusions range from a few centimetres to 10 m across. Rounded and irregular inclusions that have diffuse margins (10-15 mm wide) appear to have reacted with the host rock and to have deformed in a softened state. Broad areas of mixed amphibolite and granite occur along some contact zones.

Sunset Lake subarea

The densest swarm of mafic dykes in these belts occurs along the northeast side of the Beaulieu River volcanic belt (Fig. 3). This is the only area where a complete transect across a swarm has been studied in sufficient detail to document its precise internal constitution, to determine relationships with volcanic belts and adjacent granitic terrane, and to compare it with sheeted dykes of well known ophiolite complexes. Helmstaedt et al. (1986), have made this analogy for a dyke swarm in the Yellowknife greenstone belt.

This north-northwesterly trending swarm, which is about 2.5 km wide and at least 20 km long, tapers out southward but continues northward beyond the limits of Figure 2 (inset map). The dyke swarm comprises a multitude of diabasic to gabbroic dykes metamorphosed to amphibolite grade, separated by screens of grey weathering, sheared and foliated

granodiorite to granite gneiss, volcanic flows, and minor dykes of felsite, pink granite and fresh diabase. Figure 35 shows a detailed transect across this swarm.

Boundary relationships. A north-northwesterly trending shear zone separates the dyke swarm from the volcanic belt to the west, and the Meander Lake Plutonic Suite bounds the swarm to the east.

Plutons of the Meander Lake suite cut obliquely across the northwesterly trend of the dyke swarm and across internal contacts of multiple dyke intrusions, and includes large (ca. 50-200 m) screens of amphibolite dykes and associated granite gneiss. This relationship is complicated by rare younger mafic dykes, not distinguished from amphibolite dykes in Figure 35, that intrude Meander Lake plutons. These are fresh dark grey to rusty brown weathering dykes that cut cleanly through the granite and do not have foliated contacts. These may be related to presumed Proterozoic dykes known throughout this area (Henderson, 1985). In Figure 2 the eastern contact of the dyke swarm is drawn arbitrarily where the rock is dominantly granitic.

The shear zone along the western side of the dyke swarm is a 100-200 m wide belt of intensely strained amphibolite in which the identity of the parent lithology (i.e. dyke or mafic volcanic rock) cannot be recognized. The amphibolite has fine, penetrative, steep (85-90°) foliation and slaty cleavage commonly containing quartz veins. Horizontal surfaces show abundant evidence for dextral sense of movement (asymmetric folds, rotated clasts, c-s band structure — White, et al., 1980; Simpson and Schmid, 1983; Ramsay and Huber, 1978).

The first recognizable parent lithologies northeast of the shear zone are invariably dyke rocks with inclusions or screens of granitic gneiss (the dyke-granitoid complex), and southwest of the zone are pillow lavas containing dykes (volcanic-dyke terrane). The most intense strain seems to be localized in the dyke-granitoid complex. No dykes have been traced across this shear zone.

High strain persists outside of the shear zone where rocks are strongly foliated for up to 400 m into the dyke-granitoid complex and for at least 300 m into the volcanic-dyke terrane.

The dyke-granitoid complex within about 600 m of the shear zone is distinctly different from the complex farther northeast. Granitic screens have been deformed and are generally present only as isolated pods and slivers within massive to strongly foliated amphibolite dykes. Southeast of Step'n-duck Lake a former granitic screen has been boudinaged, broken, and strung out into a train of inclusions aligned within the plane of foliation. Foliation in the amphibolite dykes is nearly vertical and trends north to northwesterly. This shearing has caused grain size reduction and locally produced fine grained schistose rocks from medium- to coarse-grained amphibolite. Locally, the foliation, granitic screens and internal chill contacts of multiply intruded dykes are folded and their axial traces parallel the main shear zone.

To the northeast of this zone of shearing (more than 400 m away from the zone) amphibolite dykes tend to be unfoliated, massive rocks that only show development of foliation at contacts with the granitic screens and along occa-

sionally discrete shear zones within dykes. However, even massive looking amphibolite dykes experienced some strain as indicated by occasional folded aplitic or quartz veins in which a dextral sense of movement is common.

In the volcanic-dyke terrane adjacent to the shear zone, pillows are highly flattened, elongated bodies containing penetrative foliation and local zones of intense shearing parallel to the shear zone. Amphibolite dykes are rare where the volcanic succession comprises andesite to dacite flows (west central part of Fig. 35) but are abundant where flows are basaltic (southern corner of Fig. 35).

Dykes. Size, distribution and contact relations. Widths of dykes range from a few centimetres to 35 m (the largest single dyke) and average dykes are less than 10-15 m wide. Most dykes are in fact multiple intrusions as indicated by numerous internal chill contacts and small dykes that intrude larger dykes. The largest multiple intrusion is about 300 m wide but most zones are less than 100 m wide. Both sides of all dykes can be accounted for within zones of multiple injection where outcrop exposure is continuous. Hence the concept of chilling bias across a dyke swarm as applied to sheeted dykes of ophiolite complexes (Kidd and Cann, 1974) is not relevant here.

Because of lithological similarity, individual dykes generally are difficult to trace with certainty for long distances. Multiple intrusion groups, however, between distinctive large granitic screens, have been traced for about 1 km, and distribution of rare but distinct lithologies, such as plagioclase-phyric rocks, along trend of dykes for distances of 2 km; this suggests that dykes may have considerable continuity.

Dykes have a strong northwesterly trend (Fig. 35, rose diagram) and contacts are dominantly vertical, rarely in ranges of 65-85°, and only one small dyke has a shallow (21°) dip.

Within a particular multiple intrusion, internal dyke-dyke contacts of large dykes are regular and trend parallel to one another, whereas smaller (centimetre to metre sized dykes) commonly undulate or weave through the intrusion, but in the same general trend as the swarm. No distinct set of dykes cuts across the main swarm.

In spite of the amphibolite grade of metamorphism, relict primary textures, patterns of changing grain size and chilled contacts are distinguished, with variable degrees of difficulty, so that the multiplicity of intrusions has been documented for a large section across the swarm. Large dykes have coarse- to medium-grained centres, fine grain size within a metre of the contact, and chilled selvages (very fine grained to aphanitic amphibolite) against older dykes or country rock. Where dykes were emplaced side-by-side, contacts are doubly chilled. Chill relationships, however, may vary along strike from doubly chilled to one-sided chill against a very fine grained adjacent dyke margin. Amphibolite dykes invariably chill against screens of granitic gneiss and volcanics.

Dyke lithology and relative chronology. Six main types of dykes are represented in the dyke swarm: (1) aphyric fine- to coarse-grained amphibolite, (2) plagioclase-phyric amphibolite, (3) aphanitic dark green and grey amphibolite,

(4) rusty brown weathering diabase, (5) felsite, and (6) pink granite.

Most dykes in the map area belong to the first type, aphyric dark greenish-grey weathering, fine-to medium-grained and occasionally coarse grained amphibolite. Weathered surfaces accentuate relict gabbroic textures in medium- and coarse-grained rocks and diabasic textures in fine grained rocks. Sheared dykes are schistose. The dykes have been completely recrystallized to an assemblage of hornblende-plagioclase-quartz-epidote and calcite.

An uncommon but distinctive lithology amongst the amphibolite swarm is plagioclase-phyric dykes that appear intermittently about 400 and 1100 m east of the shear zone. Dykes are 3 to 15 m wide and contain euhedral to subhedral plagioclase phenocrysts (saussuritized), 5 to 50 mm across, and glomeroporphyritic clusters that make up 15-30 % of the rock. Phenocrysts are concentrated in some dyke centres.

Dark green very fine grained to aphanitic dykes and dykelets are a late phase of the amphibolite dykes. These are narrow dykes, generally 10-30 cm wide, that undulate across or along internal boundaries of larger dykes locally bifurcating and sprouting irregular apophyses.

Felsic dykes are rare in the swarm. They are nonfoliated pale grey aphanitic felsite and massive pink fine grained granite, that intrude amphibolite dykes and granite gneiss. Some grade into fine grained pink granitic dykes similar to plutons of the Meander Lake suite, whereas others are small bodies isolated in the amphibolite swarm.

Except for what can be determined in local multiple intrusion units and the late dark green dykes, no general sequence of emplacement of the amphibolite dykes within the swarm has been determined. Consistent northwesterly trend of the vast majority of dykes and the numerous subparallel internal contacts suggest that they represent a series of penecontemporaneous injections.

Discussion. Detailed mapping of a single transect across this dyke swarm suggests that the swarm probably represents a major period of multiple injection of voluminous mafic magma into granitic gneiss that had a previous history of deformation. If at least some of the mafic magma represented by this dyke swarm erupted at the surface, then it is likely that the granitic gneiss through which the dykes intruded may be hitherto unrecognized Archean basement (no radiometric dates have yet been obtained from this unit). Although most dykes are multiple intrusions, unlike sheeted dykes of ophiolite complexes there is no extensive or continuous areas of 100 % dykes and no transition to plutonic gabbros: granitic screens are abundant throughout the dyke complex on the east side of the shear zone and volcanic screens on the west side.

Density of dykes here, as in the swarm at the Cameron River volcanic belt, is not much greater than that of some high standing volcanic complexes, such as in Hawaii (Walker, 1986). The dyke swarm represents 1800 m of extension over a distance of 3000 m in a granitic gneiss terrane (i.e. 60 % extension).

The shear zone may be a major tectonic boundary at which the volcanic belt and dyke-granitoid terrane have been tectonically juxtaposed: a situation similar to that proposed along the east side of the Cameron River volcanic belt (Kusky, 1986a, c, d, in press). The dyke-granite complex may be a tectonic slice of continental crust and its contained dyke swarm and the dykes of the main swarm may not be related to the adjacent Beaulieu River volcanic suite nor the dykes it contains.

It has not been demonstrated that the dyke swarm in the Sleepy Dragon Complex is the same as that in the Cameron River volcanic belt or that they are related to volcanism that formed the belt. The mafic dyke swarms record extension in the basement gneisses and in the substratum upon which the volcanic belt was deposited (not necessarily at the same place). Even if dykes in the Sleepy Dragon Complex are not directly related to the adjacent volcanic belt, they record an important mafic magmatic event that likely resulted in effusion of lava on the surface of sialic (continental) crust. Thus, the Sleepy Dragon Complex probably was basement to some volcanism.

Proterozoic mafic dykes

Proterozoic mafic dykes of various trends in the Yellowknife-Hearne map areas generally are fresh, undeformed units that have been assigned to various sets mainly on the basis of their orientation (Henderson, 1985). These dykes are sparse in the present map area and are not subdivided in Figures 2 and 3.

Northwesterly trending dykes of fresh diabase and gabbro, generally less than 50 m wide, cut across the Cameron River belt at high angles. These dykes stretch discontinuously for some 20 km with fairly constant trends through the Sleepy Dragon Complex, volcanic belts and the Burwash sediments near Cameron River (Henderson, 1938, 1941) and northwesterly trend suggests that they may belong to the Indin dykes (McGlynn and Irvine, 1975; Burwash et al., 1963).

Two northwesterly trending dykes follow or cut the volcanic stratigraphy at low angles in the Sunset Lake subarea. One north of Sunset Lake is a hornblende diabase that is continuous for 4 km. Numerous minor displacements of this dyke reveal a multitude of small faults in the volcanic terrane that otherwise would be unrecognizable. The other is a pyroxene gabbro dyke south of Payne Lake.

These rocks are rusty-brown weathering, massive diabase and gabbro that are unusually fresh compared to other mafic intrusions within the Beaulieu Group. The dyke south of Payne Lake is quartz-bearing pigeonite diabase that has a well developed ophitic texture (Fig. 37). Mafic minerals are pigeonite and augite that have thin rims of green-brown to blue-green hornblende. The hornblende diabase dyke north of Sunset Lake contains brown to blue-green (on rims) hornblende that rarely preserved patches of clinopyroxene in their cores. This hornblende is not poikiloblastic. Hornblende in this rock is interpreted as a late stage reaction where most of the pyroxene has been converted to amphibole. Euhedral to subhedral laths of plagioclase have normal to oscillatory



Figure 37. Photomicrograph of fresh quartz-bearing pigeonite diabase from post-Archean dyke cutting the Sunset Lake belt south of Payne Lake. Scale bar 2 mm long. GSC 190982

zoning with cores of labradorite and rims of andesine. Quartz forms anhedral grains or granophyre that occupies angular areas between plagioclase and pyroxene.

These dykes are younger than the Yellowknife Supergroup, have not been involved in the metamorphism and deformation that affected the volcanic belts, and are considered to be part of the Mackenzie dyke swarm.

Dykes of the Milt diabase (Henderson, 1985) differ from other Proterozoic dykes in that they form gently dipping to subhorizontal sheets that in most cases occur singly. One of these dykes forms a sinuous pattern within the Sleepy Dragon Complex between Detour and Turnback lakes. Others form thin sheets in the Cameron River Basalt that are exposed in a northwesterly trending cliff face west of Patterson Lake.

Felsite dykes

Felsite dykes, not related to rhyolitic formations, cut most map units in all three volcanic belts. In the Tumpline Lake subarea aplite, aphanitic felsite and granitic dykes intrude the volcanic belt, the Burwash Formation and granitic plutons and gneisses. North of Cleft Lake, a swarm of aplite and granitic dykes, 0.25 to 6 m wide, trends north-northwesterly through the volcanic and plutonic rocks. In the Cameron River belt, aplite dykes cut amphibolite dykes and granitic rocks east of Webb Lake. In the Sunset Lake

belt, felsite dykes intrude large amphibolite sills west of Amacher Lake near the contact with granitic terrane to the west.

These felsite dykes are clearly younger than the volcanism, major regional deformation and some plutons. They are presumed to be related to post-deformational granitic plutons.

METAMORPHISM

Regional metamorphism

The Cameron River and Beaulieu River volcanic belts are regionally metamorphosed in the amphibolite facies, except for a north-northwesterly trending zone, about 4 km wide, of greenschist facies that follows the valley of Sunset Lake (Fig. 38). The boundary between the greenschist and amphibolite facies is drawn at the cordierite isograd in metasediments (as defined by Henderson, 1976) and at the first appearance of blue-green hornblende in volcanic rocks of mafic to intermediate compositions. Amphibolites in which actinolite is the dominant amphibole are considered to be in the upper greenschist facies.

Sediments of the Burwash Formation to the south and west of the volcanic belts are chiefly in the greenschist facies. In most places the transition from greenschist to amphibolite facies occurs near the boundary between the Burwash Formation and volcanics of the Beaulieu Group. The cordierite isograd in sediments, however, also appears to be spatially related to some plutonic bodies.

Textures and mineralogy of metamorphosed volcanic rocks

In passing from greenschist to amphibolite facies, rocks within the volcanic belts go through distinctive changes in colour, degree of preservation of primary textures, and development of metamorphic textures and mineralogy (Table 2). Textural and mineralogical contrasts between rocks in greenschist and amphibolite facies are summarized below for the main lithologies in the volcanic belts.

Metarhyolites and metadacites

In the greenschist facies of metamorphism primary textures of felsic volcanic rocks (rhyolites and dacites) are well preserved and show only minor evidence of recrystallization. These textures include euhedral, subhedral and embayed phenocrysts, feldspar microlites and delicate features such as cuspidate shards, pumiceous textures and spherulites.

Phenocrysts of quartz and plagioclase have sharp boundaries that do not appear to have reacted with matrix minerals and are optically continuous crystals without recrystallization or polygonization at their margins (except for those that have been affected by high strain). Boundaries of plagioclase microlites, however, are minutely intergrown with quartz and feldspar of the matrix but the microlites are not polygonized.

Texture of the matrix varies from spherulitic (showing no evidence of recrystallization) to extremely fine (10-20

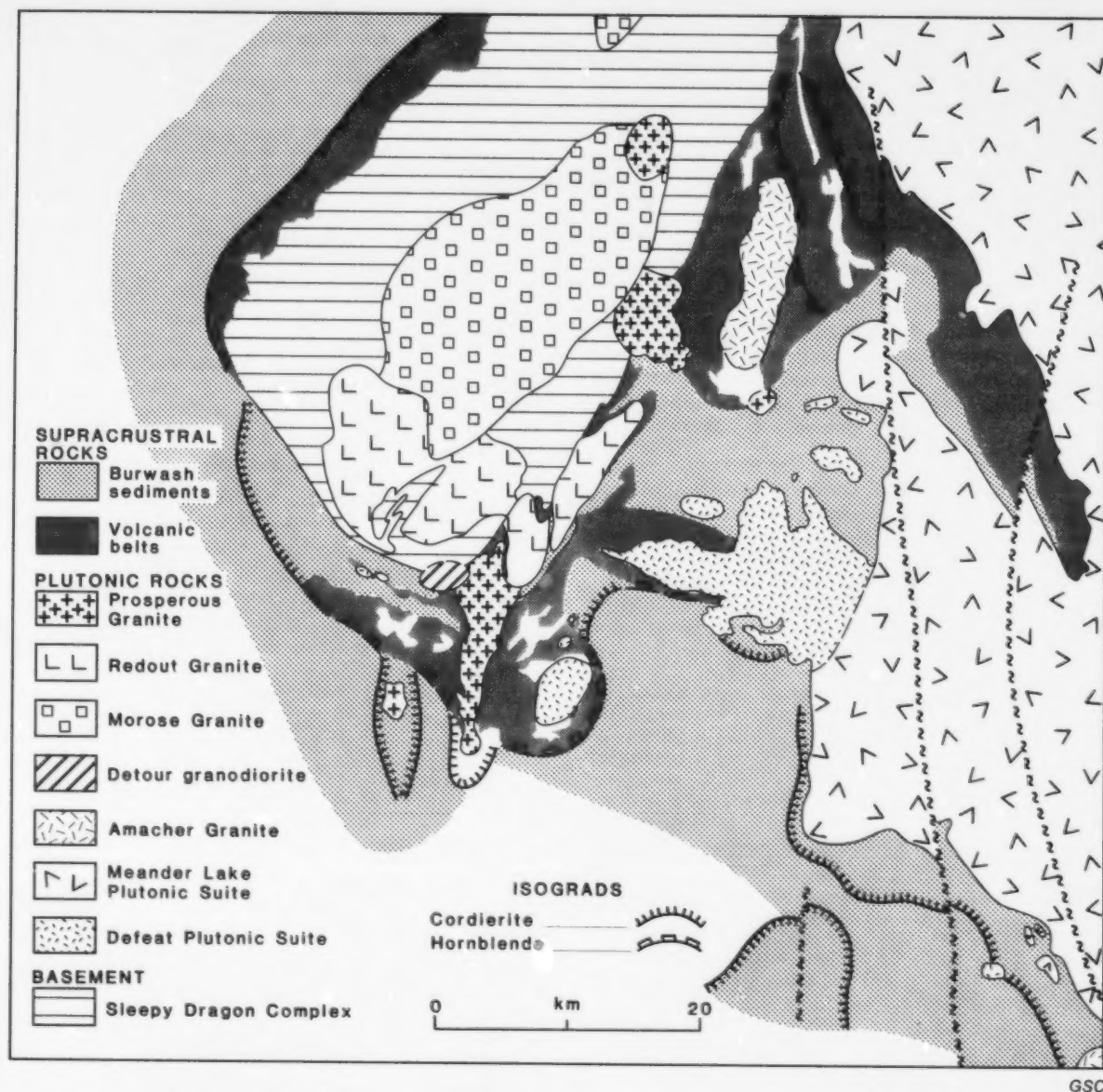


Figure 38. Distribution of granitic plutons, basement and metamorphic isograds near the Cameron River and Beaulieu River volcanic belts. Cordierite isograd and plutons within the Sleepy Dragon Complex and south and east of the volcanic belts, after Henderson (1985).

micron grains) microfelsic aggregates of quartz and feldspar. The microfelsic aggregate varies from inequigranular micropoikilitic areas to irregular granoblastic areas comprising interlocking grains that have irregular or sutured boundaries. Plagioclase in the aggregates is commonly untwinned albite.

Generally, mafic minerals are sparse in rhyolites and metamorphic assemblages are:

calcite — chlorite — biotite — quartz — albite
calcite — sericite — quartz — albite

In passing from greenschist to amphibolite facies the main changes in rhyolitic rocks are in the texture of the felsic matrix and in the nature of phenocryst boundaries. Phenocrysts may be completely polygonized or partly polygonized on margins and boundaries are intergrown, with crystals forming the matrix. In spite of the recrystallization, phenocrysts retain their primary euhedral, subhedral to rounded habits.

Matrix of rhyolite in the amphibolite facies is a felsic aggregate, generally coarser than that in rocks of greenschist facies metamorphism. Textures may vary in the same rock from micropoikilitic (comprising irregular grains ranging

Table 2. Summary of mineralogy and textures of metavolcanic and metasedimentary rocks of the Beaulieu Group and the Burwash Formation

Lithology	Greenschist facies metamorphism	Amphibolite facies metamorphism
Rhyolite	QZ-AB-CC $\pm$ CH, BT, SE Delicate primary textures preserved -shards, microlites, spherulites, clastic textures. Sharp boundaries on phenocrysts.	BT-QZ-PC $\pm$ HB, MI, GT, EP, TO, CC Microfelsic matrix- granoblastic; delicate textures lacking. Phenocrysts have irregular boundaries that are intergrown with matrix.
Mafic volcanics and dykes	Pale green, pale grey. EP-CH-QZ-AB $\pm$ AC, CC PC microlites and diabase textures well preserved Mafic minerals altered to chlorite, AC(EP)?	Medium green, dark green, black. HB-QZ-PC $\pm$ EP, BT, CC (HB blue-green) Primary PC, polygonized PC in fine aggregates is oligoclase to andesine. Microlites recrystallized to felsic aggregates and PC phenocrysts partly to completely recrystallized. Relics of gabbroic textures are recognizable.
Carbonate		DI-CC-QZ-PC $\pm$ SP, EP, MI, AC Granoblastic carbonate
Intravolcanic sediments		BT-QZ-PC $\pm$ GT, ST, MV Mica schists
Burwash sediments		CD-BT-QZ-PC $\pm$ ST, SL Mica schists
Iron-formation		QZ-MT-AC-HB Granoblastic quartz
AB = albite AC = actinolite BT = biotite CC = calcite CD = cordierite CH = chlorite	DI = diopside EP = epidote GT = garnet HB = hornblende MI = microcline MT = magnetite MV = muscovite	PC = plagioclase QZ = quartz SE = sericite SP = sphene SL = sillimanite ST = staurolite TO = tourmaline

from 5-10 microns across and undulose patches of polygonized grains up to 60 microns across) to granoblastic aggregates of equant quartz and feldspars where grain size may range from 30-80 microns. Some granoblastic aggregates have elongate crystals that have weak to strong preferred orientation in the plane of mica schistosity. Plagioclase in the matrix is oligoclase to andesine. Delicate textures, such as microlites and spherulites, are absent.

Hornblende appears in sparse amounts in some rocks or as major constituents where mafic material is interlayered with felsic volcanic material. Hornblende and garnet, where present, poikilitically enclose quartz and feldspar of the matrix.

Typical metamorphic assemblages of rhyolites and dacites in amphibolite facies include:

garnet — biotite — hornblende — quartz — plagioclase
calcite — muscovite — biotite — hornblende — quartz — plagioclase
tourmaline — calcite — sericite — biotite — quartz — plagioclase
epidote — muscovite — biotite — microcline — quartz — plagioclase

Meta-andesites and metabasalts

Basalts and andesites go through a much more drastic change in colour, mineralogy and texture than rhyolites, when passing from greenschist to amphibolite facies of metamorphism. Mafic volcanic rocks in greenschist facies are pale green to pale grey, and very fine grained to aphanitic. They preserve delicate primary textures where felsic minerals are involved.

Phenocrysts of plagioclase, having very sharp outlines, are not polygonized and show little evidence of reaction with mafic minerals. Delicate zoning is well preserved. Plagioclase microlites show feldt to trachytic fabrics. Microlites have minutely irregular boundaries and ragged terminations that appear to have reacted with surrounding metamorphic quartz and plagioclase aggregate. Plagioclase microlites in the upper greenschist facies, however, are polygonized relics of original laths in which polysynthetic twinning can still be identified. The degree of preservation of textures declines where mafic minerals are involved. No primary pyroxene is preserved.

Areas between plagioclase microlites that are filled with mixtures of chlorite, epidote, actinolite and calcite (in the upper greenschist facies) presumably represent sites of inter-

stitial pyroxene of the original rock. In the upper greenschist facies these areas are filled with minute needles of actinolite.

Mineral assemblages include:

epidote — chlorite — tremolite — quartz — feldspar
epidote — calcite — chlorite — quartz — feldspar

Basalts in amphibolite facies, in marked contrast to those in the greenschist facies, are dark green to black, very fine grained to locally medium grained rocks. Plagioclase in rocks of lower amphibolite facies has well preserved zoned cores of primary andesine to labradorite, but margins are recrystallized to form irregular boundaries and polygonized zones that are intergrown with the surrounding crystalloblastic aggregates of quartz, feldspar and mafic minerals. In higher grades of amphibolite facies metamorphism, the rocks are completely recrystallized and plagioclase is totally polygonized or reduced to micropoikilitic and granoblastic aggregates. Recrystallization has obliterated primary igneous textures.

Original mafic minerals are completely replaced by poikiloblastic-crystalloblastic hornblende. Both blue-green hornblende and actinolite coexist near the hornblende isograd (for example, on the west side of the Cameron River belt). Hornblende in some rocks is pale blue-green and commonly has blocky outlines or stubby rectangular forms, suggesting it is replaced pyroxene. Some rocks preserve ophitic textures in spite of the high degree of recrystallization of mafic minerals. Hornblende in higher grades is deep blue-green to green and intensely pleochroic, in contrast to the pale blue-green in the lower amphibolite facies. Metamorphic assemblages of basalts in the amphibolite facies are:

sphene — epidote — hornblende — quartz — plagioclase
hornblende — quartz — plagioclase.

Amphibolite dykes

Amphibolite dykes within volcanic belts are completely recrystallized. Plagioclase is polygonized to fine, irregular crystalloblastic aggregates or poikiloblastic aggregates in which relict twinning can still be identified. Amphibole is typically blue-green hornblende that poikiloblastically encloses minute grains of plagioclase, quartz, epidote and calcite. In spite of the high degree of crystallization, relics of primary, gabbroic or diabasic textures are preserved.

Dykes within granitic terrane east of the Cameron River belt are less intensely recrystallized and appear to have been shielded from the metamorphism that affected the adjacent volcanic belt. The plagioclase in these rocks is not polygonized and retains delicate oscillatory zoning. Mafic minerals, however, are completely recrystallized to blue-green hornblende. Hornblende that has sharp outlines and is not poikiloblastic appears to have replaced pyroxene grain-for-grain.

Metavolcaniclastic rocks, sediments and carbonates

Volcaniclastic rocks show similar mineralogical changes as massive equivalents in passing from greenschist to amphibolite

lite grades, except that these rocks tend to be more aluminous, as suggested by the abundance of biotite, the common occurrence of garnet, and locally, staurolite. Fine clastic textures are preserved in greenschist facies but are generally unrecognizable in amphibolite facies. In amphibolite grades only coarse clastic textures and bedding are preserved.

Contrasting metamorphic assemblages in adjacent layers of thin-bedded rocks may reflect a primary compositional layering. For example, thin beds in the Payne Lake Formation are sharply defined by distinct assemblages of mafic minerals in each layer: (1) hornblende only, (2) hornblende-actinolite, (3) actinolite-biotite, or (4) garnet-actinolite-biotite.

Intravolcanic sediments south of Payne Lake, metamorphosed to amphibolite grade, generally lack hornblende, like sediments of the Burwash Formation. These rocks are mica schists that have assemblages: (1) biotite-quartz-feldspar, (2) garnet-biotite-quartz-feldspar, and (3) biotite-muscovite-garnet-staurolite-quartz-feldspar.

Iron-formations in amphibolite grades are completely recrystallized laminated rocks comprising fine to coarse granoblastic quartz and variable amounts of actinolitic hornblende, and magnetite, with minor amounts of epidote.

Carbonate units, which invariably contain abundant felsic volcanic material, are all in the amphibolite facies and have metamorphic assemblages: (1) actinolite-diopside-microcline-quartz-plagioclase-calcite (dolomite); (2) epidote-sphene-calcite; and (3) quartz-calcite-diopside-hornblende-vesuvianite (Goetz, 1974).

Summary

The following changes take place with prograde metamorphism of mafic volcanic rocks in the Cameron-Beaulieu volcanic belts. Amphibole changes from actinolite to pale blue-green hornblende to deep blue-green hornblende to green hornblende. Plagioclase changes from albite to oligoclase and andesine. Chlorite and actinolite gradually disappear in favour of biotite and hornblende. Experimental work of Moody et al. (1983) documents similar changes across the greenschist-amphibolite boundary in mafic systems. Fine grained rocks become coarse grained and quartz, calcite and epidote locally occur as segregations. Primary textures become progressively obliterated. Phenocrysts become more polygonized and intergrown with metamorphic minerals of the matrix. Table 2 summarizes mineralogical and textural changes of the various lithologies in the volcanic belts.

Several authors have documented the change in pleochroism of amphiboles with increasing grade of metamorphism and the changing chemistry. Dekker (1978) has correlated pleochroic colours of calcic amphiboles with chemical composition and found that (1) colour of amphibole is mainly determined by Ti and Fe⁺³ content, (2) brown amphiboles, on an average, are richer in Ti, alkalis and Fe-total, whereas green calcic amphiboles are richer in manganese, silica and Fe⁺³. Dekker (1978), Vejnar (1977) and Miyashiro (1968) have shown that amphiboles change from blue-green to green to brown with increasing metamorphic grade from amphibolite to granulite facies. Actinolitic hornblendes are typical of greenschist facies. Maruyama et al. (1983) have

documented mineral parageneses and systematic changes in mineral composition for the greenschist-amphibolite transition in basaltic rocks from the Yap Islands and Japan under low pressure conditions.

Contact metamorphism

Contact metamorphism in the volcanic belts appears to be limited to narrow aureoles around granitic plutons and immediately adjacent to some amphibolite dykes. A 20 m wide contact metamorphic aureole occurs within the Burwash sediments along the southern and western sides of the trondhjemite pluton at the south side of Detour Lake. The aureole is characterized by a slight increase in grain size (noticeably of biotite) in the sediments and by a network of quartz veinlets and lenses less than 30 cm long.

A contact metamorphic aureole around the pluton of Redout Granite northwest of Turnback Lake is recognized by (1) a slight coarsening of grain size and enrichment of muscovite in felsic volcanics that produces a sugary texture and schistose structure in the rocks; (2) mica-enrichment in greywackes and shales, in the appearance of garnet-cordierite-biotite assemblages in pelitic rocks along the eastern side of the pluton; and (3) very coarse grained, garnet-diopside skarn zones (7 m wide) adjacent to small intrusions into carbonate rocks off the northeastern end of the pluton. Silicified metasediments and metavolcanics near the northwestern corner of Turnback Lake contain pyrite, chalcopyrite and magnetite mineralization.

Contact metamorphism around the Prosperous Granite pluton west of Tumpline Lake is defined by the presence of abundant muscovite and microcline in felsic volcanic rocks at the contact with the pluton.

Contact metamorphism related to amphibolite intrusions has been inferred in a few localities. Sediments in felsic volcanics adjacent to the large amphibolite intrusion along the north side of Tumpline Lake are characterized by development of unusually coarse cordierite, garnet and mica. Metarhyolites adjacent to the western side of the dyke are coarse grained, garnet-biotite-muscovite schists and some metasediments are coarse grained cordierite-mica schists containing cordierite crystals up to 10 cm long and 2 cm wide.

In some localities in the Cameron River belt, the grain size of amphibolite pillow lavas becomes coarser at the immediate contact with some large amphibolite sills. This textural change is interpreted to be related to contact metamorphism.

Discussion

A model explaining distribution of regional metamorphism in the area mapped should account for (1) the apparent change from greenschist to amphibolite facies at boundaries between sediments of the Burwash Formation and volcanics of the Beaulieu Group; (2) the band of greenschist metamorphism down the centre of the Sunset Lake belt, (3) the cordierite isograd locally related to plutons in some areas but not in others; (4) the timing of metamorphism with respect to deformation, and (5) the environment in which metamorphism took place.

Previous investigators have concluded that metamorphism climaxed near the end of, or after, the period of folding that produced the dominant fold patterns in the Yellowknife Supergroup. Fyson (1975, p. 771) in a structural study of the Burwash sediments south and west of the Cameron River and Beaulieu River belts, concluded that "growth history of the various metamorphic minerals from pre- to post-D₃ deformation suggests that they reflect progressive regional metamorphism which culminated with emplacement of late granite plutons". Thompson (1978) who made a regional assessment of metamorphism in the Slave Province, suggested (1) that the peak of metamorphism occurred after the main phases of folding, and (2) that cordierite-biotite-actinolite-sillimanite (staurolite) assemblages in the metasediments are typical of those produced by low pressure regional metamorphism of Pyrenean-type. He attributed the patchy pattern of greenschist and amphibolite grade metamorphism to differential erosion of an irregular thermal topography (thermal ridges or domes and depressions) preserved in the rocks as an isograd pattern. The metamorphism has been superimposed on the rocks of the Yellowknife Supergroup during a tectonic event about 2600 Ma ago.

Henderson (1985) suggested that the pattern of metamorphism can be related to emplacement of granitic plutons although he considered the width of the aureoles too broad to be a result of contact metamorphism in a strict sense. The relationship seems obvious in the Tumpline Lake subarea, and may be a plausible explanation in the Sunset Lake belt, where amphibolite grade metamorphism occurs within 3 km of granitic intrusives. The argument is less convincing in the Cameron River subarea where the amphibolite metamorphism is about 10 km from the nearest pluton (Morose Granite) and mafic dykes are more altered in the volcanic belt than in the Sleepy Dragon Complex, which is closer to the pluton.

Although amphibolite facies are indicated by (1) the cordierite isograd in Burwash sediments, and (2) the hornblende isograd in volcanics, the two isograds do not necessarily represent the same temperature and pressure conditions and may not be coincident. Hornblende in mafic volcanics may be stable at temperature and pressure conditions slightly lower than cordierite in the sediments. Thus the apparent sudden jump into amphibolite facies metamorphism at the contact between volcanic sediments may be in part a function of composition.

Fonteilles (1968) and Autran et al. (1970) have documented convergence of isograds in Paleozoic cover rocks around basement cores in the Pyrenees. They considered this phenomena as being related to dehydration of cover rocks during metamorphism. Dehydration is a strongly endothermic reaction. Thus, more energy is required to convert a hydrous rock to the same grade of metamorphism as an anhydrous rock. In general, basement rocks that have suffered high grade metamorphism are anhydrous, whereas cover rocks (sediments at low grades of metamorphism) are abundantly hydrous. Isograds would rise at a higher rate in the anhydrous basement than in the hydrous cover rocks because of the energy required to affect dehydration. The net effect is a steepening of isograds at the contact between basement and cover rocks. The same phenomena may apply to a contrast in the composition between mafic volcanic rocks of the Beaulieu Group and sediments of the Burwash Formation (basalts are relatively anhydrous compared to sediments).

The compositional contrast between the two lithologies, as well as the effect of the granitic basement hot horsts in the Beaulieu River and Cameron River belts, may account for the depression of isotherms and the resultant abrupt change in metamorphic grade at the boundary between Burwash sediments and the volcanic belts.

Experimental studies of phase relations between greenschist and amphibolite assemblages in rocks of basaltic composition by Liou et al. (1974) indicated that the upper boundary of the greenschist assemblage albite-chlorite-epidote-actinolite is a 475°C and the lower boundary of the amphibolite assemblage plagioclase-hornblende is at 550°C at 2 kb (2×10^5 kPa) fluid pressure. They also noted that increased fugacity of oxygen would reduce significantly the temperature range of the transition. Maruyama et al. (1983), however, supported the concept of a transition loop between actinolite and hornblende and that a polymorphic transition of actinolite takes place at 420°C at 2 kb (2×10^5 kPa), and about 450°C at 3 kb (3×10^5 kPa) fluid pressure.

Various authors have speculated on the conditions of metamorphism in this part of the Slave Province based on mineral assemblages in metasediments. Lack of kyanite amongst the aluminosilicates suggests relatively low pressures. Ramsay and Kamineni (1975) and Kamineni et al. (1979) suggested that metamorphism in areas between Yellowknife and Ross Lake may have involved several stages in which pressure and temperature conditions varied. They suggested that pressures during early stages ranged between 3 and 4.5 kb (3×10^5 and 4.5×10^5 kPa), with temperatures up to 500°C, but dropped to 2.5-3.5 kb ($2.5-3.5 \times 10^5$ kPa) during a later stage. Thompson (1978) suggested that the mineral parageneses described by Ramsay and Kamineni can develop entirely at lower pressures.

PETROCHEMISTRY

All rocks in these volcanic belts are variably altered because of metamorphism that accompanied extreme tectonic deformation as well as all the other deuteric, hydrothermal, and diagenetic changes that are known to be inflicted upon volcanic rocks after deposition. In spite of this inherent alteration, major stratigraphic units within each volcanic belt have distinctive chemistries and appear to have retained some degree of chemical integrity. By analogy with modern volcanic rocks, the compositions of the volcanic units within each belt evolved with time from mafic tholeiites to felsic calc-alkaline rocks.

Sampling and analysis

One hundred and seventeen samples from the three volcanic belts were analyzed for 14 major elements and 15 minor and trace elements. Appendix 2 tabulates volcanic formations, chemical analyses, normative compositions and specific gravity for each sample. Locations of samples are shown in Figure 8 for the Cameron River subarea and index maps for each formation in the Tumpline and Sunset Lake belts (Fig. 12, 13, 21, 26) as well as in Figures 2 and 3.

At each location a composite sample was taken from an outcrop area 1 to 2 m² by collecting 0.5 to 2 kg of chips

of the "freshest rock" considered representative of that area. "Fresh" rock is that material that is free of (1) altered weathering rinds, (2) intense fractures lined with alteration minerals, (3) abundant carbonate (does not react with HCl) and (4) veins of carbonate or silica. A hand specimen representative of the rock sampled was taken for thin section study and specific gravity measurements. For many sample locations this procedure was repeated in two or more areas in order to estimate chemical variability on the scale of the outcrop.

Chemical analyses were performed in the laboratories of the Geological Survey of Canada between 1974 and 1975 for samples collected by me, and in 1964 and 1967 for samples collected by W. A. Baragar. Major elements were analyzed by X-ray fluorescence spectroscopy* for Si, Al, Mg, Fe (total), Mn, Ca, K, and Ti and by rapid chemical methods for Na, P, Fe⁺², H₂O and CO₂. Trace elements and minor elements were analyzed by quantitative spectrographic methods. The accuracy of the major element values expected by the laboratory is as follows:

SiO ₂ % (0.3 + 1 %),	Al ₂ O ₃ % (0.3 + 1 %),
Fe ₂ O ₃ % (0.6 + 1 %),	FeO % (0.3 + 2 %),
MnO % (0.01 + 1 %),	MgO % (0.16 + 1 %),
CaO % (0.05 + 1 %),	Na ₂ O % (0.1 + 2 %),
K ₂ O % (0.03 + 1 %),	TiO ₂ % (0.01 + 1 %),
P ₂ O ₅ % (0.01 + 1 %),	CO ₂ % (0.02 + 3 %),
H ₂ O % (0.05 + 5 %),	S % (0.02 + 5 %).

Spectrographic analyses are accurate to within 15 % of the value reported for Sr, Ba, Cr, Zr, V, Ni, Cu, Y, Co, Sc and within 30 % of the value reported for Zn, Pb, Ga, Sn, and Ag.

XRF analyses were compared with analyses by classical methods for one sample of rhyolite and four samples of basalts. In Appendix 2, samples are designated by a subscript "c" if analyzed by classical methods and "x" if analyzed by XRF methods. Classical analyses tend to give slightly higher values for SiO₂, Al₂O₃, and alkalis, whereas XRF analyses give higher values for FeO and MnO.

Alteration

The presumed multi-stage alteration presents problems in classifying the rocks since classification schemes are developed using analyses of fresh, usually Cenozoic rocks. In addition, petrogenetic interpretations are ambiguous because of the superimposed effect of alteration on signatures of petrogenetic processes.

Quantitative estimates of element mobility in relatively low grade Archean sequences (Gelinas et al., 1977, 1982; Ludden et al., 1982; Condie et al., 1977) suggest that mobility of elements Na, K, Ca, Sr, Rb, Ba, is large and significant

* X-ray fluorescence determinations are made on fused discs. Discs are prepared by mixing the sample with lithium tetraborate, lithium fluorite and ammonium nitrate, fusing in a Claisse fluxer, then pouring into a mould. The resulting disc is read by X-ray fluorescence spectrometry and data reduction made by computer.

whereas Al, Ti, P, Zr, Ni, Cr and the REE are less mobile. Considering the suspected element mobilities, least altered samples were selected following criteria modified after those suggested by Gelinas et al. (1977) as follows:

- 1) $\text{CO}_2 < 2.5$ and $\text{H}_2\text{O} + \text{CO}_2 < 5$
- 2) Normative corundum < 1
- 3) $\text{SiO}_2 > 80\%$
- 4) Samples with unusually high or low Ab or An when plotted on the Ab-An-Or diagrams are considered altered (Fig. 39).

Normative corundum is a particularly sensitive indicator of alteration of all alkali elements. Of the 134 samples analyzed, 44 meet these conditions, and are considered to be the best estimate of the original rock composition. Table 3 lists the percentage of least-altered samples for formations in each subarea. The Sunset Lake subarea has apparently undergone the most pervasive alteration with respect to the other two subareas.

Appendix 2 includes analyses of all specimens of the Cameron River and Beaulieu River volcanic belts. A dot beside sample numbers indicates rocks that are considered least altered according to the above criteria. Oxides of silica, calcium and alkalis appear to be the most affected by metasomatism, whereas iron, manganese and Al_2O_3 were altered only to a small degree. Altered rhyolites tend to be enriched in SiO_2 but depleted in CaO and Na_2O relative to unaltered rocks. Altered basalts from the Cameron River subarea show a depletion in Na_2O . Andesites (of which 50% of the samples are altered) generally tend to be enriched in SiO_2 , depleted in CaO, and possibly enriched in Na_2O .

Most altered rocks fall within fields of the unaltered rocks in the Jensen plot (Fig. 40b). Alteration has not affected the parameters of the plot (i.e. Fe, Mg, and Al) as greatly as silica, alkalis and CaO, upon which most of the other diagrams are based. Consequently, the AFM diagram (Fig. 40a) shows a close clustering of both altered and unaltered compositions for each formation.

Table 3. Proportions of least altered to total samples analyzed (L/T) in formations from the Cameron-Beaulieu volcanic belts.

Formation	L/T	%
<u>Cameron River Subarea</u>		
Rhyolite	1/3	33.3
Webb Lake Andesite	4/8	50.0
Cameron River Basalt	13/24	58.3
Maffic dykes	5/6	83.3
<u>Tumpline Lake subarea</u>		
Sharrie Rhyolite	1/2	50.0
Turnback Rhyolite	3/9	33.3
Tumpline Basalt (andesite member)	2/4	50.0
Tumpline Basalt	2/2	100.0
<u>Sunset Lake subarea</u>		
Rhyolite	0/6	00.0
Alice Formation (dacite member)	3/12	25.0
Andesite	5/21	23.6
Basalt	4/25	16.0

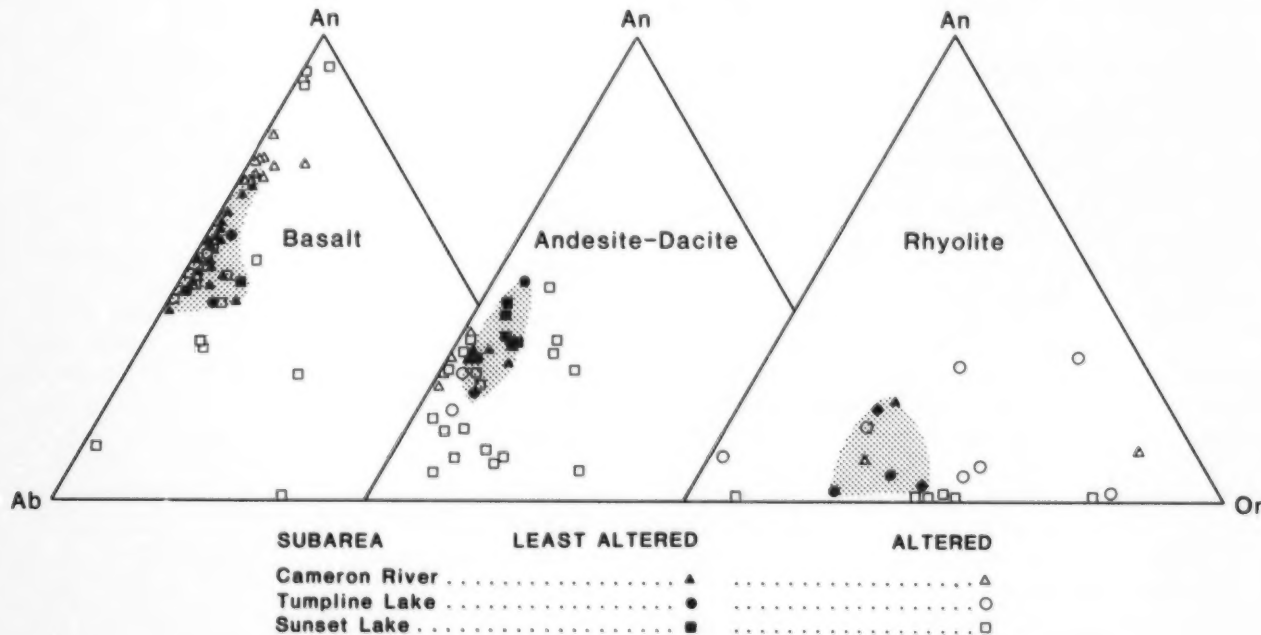


Figure 39. Least altered (grey areas and solid symbols) and altered (open symbols) rocks from the Cameron-Beaulieu volcanic belts plotted on Ab-Or-An diagram.

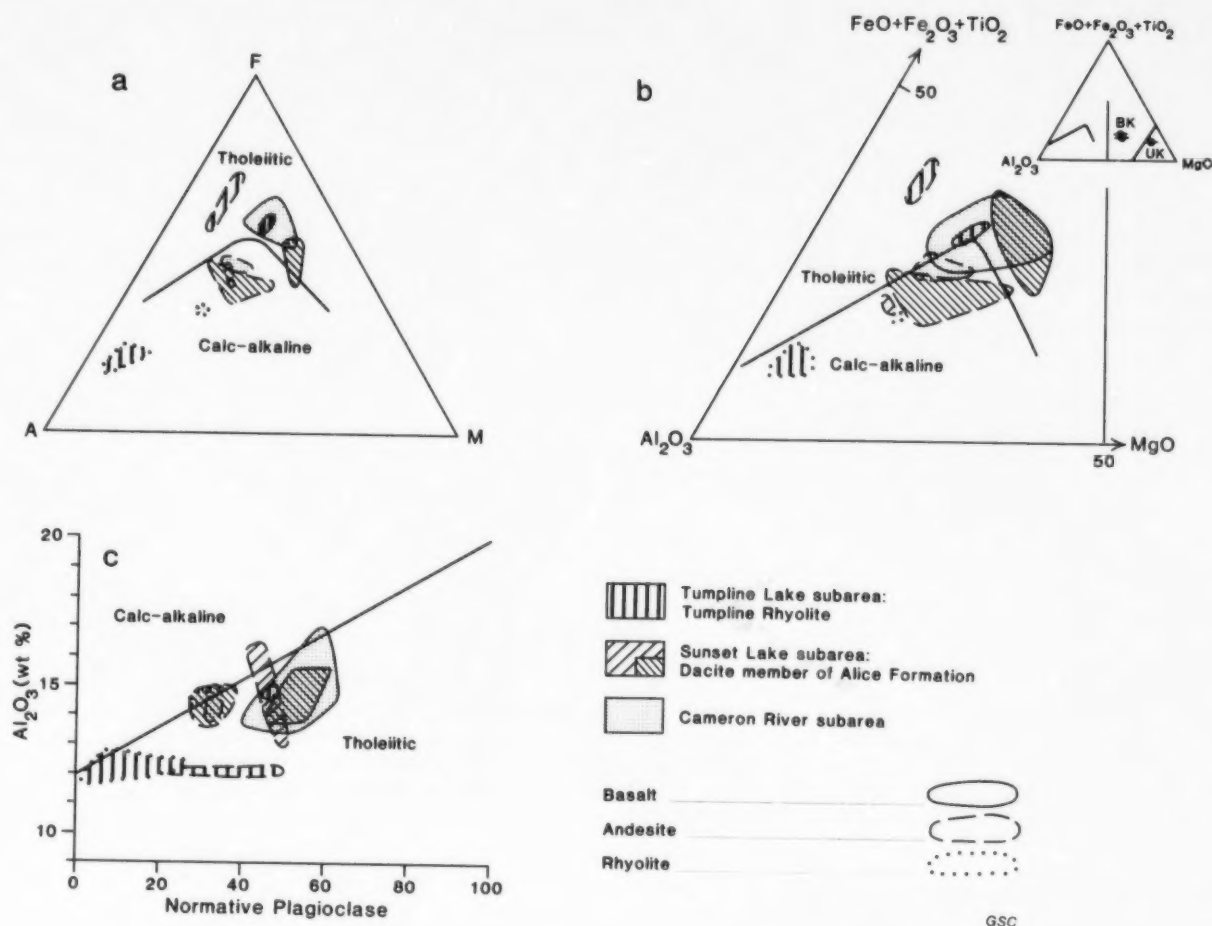


Figure 40. Volcanic formations of the Cameron-Beaulieu volcanic belts plotted on (a) AFM diagram, (b) Jensen cation plot (Jensen, 1976) and (c) Al_2O_3 -normative plagioclase diagram (Irvine and Baragar, 1971). These diagrams show the mixed tholeiitic and calc-alkaline affinities for volcanic rocks in the various subareas. BK = Basaltic komatiite, UK = Ultra-mafic komatiite.

Geochemical classification

In the following discussion, conclusions regarding classification are based on least altered samples. Unless otherwise noted, most classification diagrams show least altered samples only. Appendix 2D contains diagrams showing all samples analyzed (both altered and least altered) from these belts.

Classification of the metavolcanic rocks used in this bulletin (Fig. 41) follows that of Irvine and Baragar (1971).

All formations in the Cameron and Beaulieu volcanic belts are subalkaline (Fig. 42). None of the samples contain normative nepheline or acmite. All least altered rocks plot within subalkaline fields of the alkalis-silica (Fig. 42e) and $\text{O1} - \text{Ne} - \text{Q}$ diagrams (Fig. 42c) using dividing lines proposed by Irvine and Baragar (1971). A small number of altered rocks, however, plot within the alkaline fields of both diagrams (see Appendix 2D). Discrimination diagrams (Fig. 42a,b,d) utilizing minor and trace elements considered by

Winchester and Floyd, (1976, 1977) to be immobile or chemically stable during metamorphism and alteration also support the subalkaline nature. Some samples that would fall in the alkaline fields in the plots of Ga vs Zr/TiO_2 and P_2O_5 vs Zr (not the same samples in each case) are rocks that are altered with respect to alkalis. Possibly the minor elements are not as immobile as the authors claim.

Rocks of the Cameron River and Beaulieu River volcanic belts fall into both the tholeiitic and calc-alkaline series (Fig. 40, 42). In all belts voluminous effusions of basalt give way to andesites, and volcanism ends with dacites and rhyolites. Mean compositions of all basaltic formations and all the mafic dykes plot within the tholeiitic fields of all diagrams. In both the Cameron River and Sunset Lake subareas tholeiitic basalts give way to calc-alkaline andesites, whereas in the Tumpline Lake subarea both basalts and andesites are tholeiitic. Fields of both basalts and andesites of the Sunset Lake subarea overlap with each other and with the boundary between tholeiitic

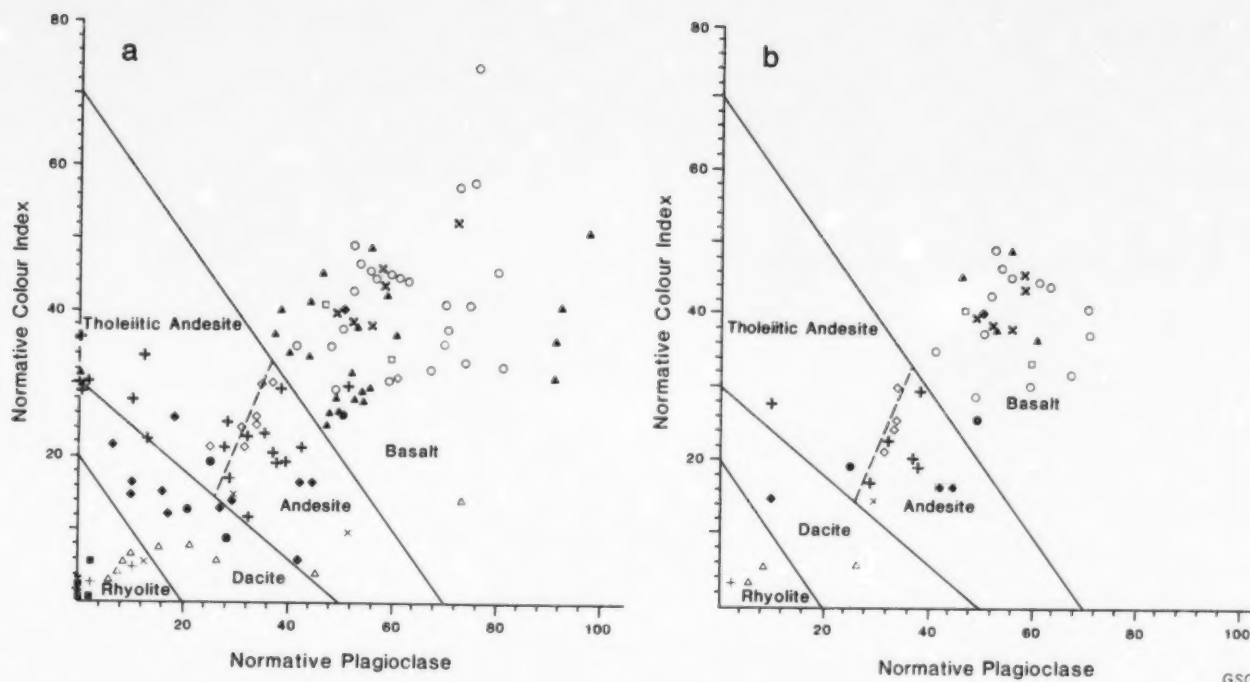


Figure 41. Composition of volcanic rocks of the Cameron and Beaulieu volcanic belts (classification after Irvine and Baragar, 1971). Normative colour index = $Ol + Opx + Cpx + Mt + Il + Ht$. Normative plagioclase composition = $100 An / (Ab + An + 5/3 Ne)$. (a) All rocks of the Cameron-Beaulieu belt analyzed. (b) Least altered samples only. Plot symbols indicate formations listed in Appendix 2.

and calc-alkaline series, suggesting that some of the magma may have had a composition transitional between the two types. Only 15 % of all the basalts analyzed are high alumina basalts and 10 % of all andesites are tholeiitic, and most of these are from the Sunset Lake subarea.

Chemical variations within the volcanic belts

SiO₂ variation diagrams

SiO₂ in least altered rocks ranges from 45-58 % in basaltic formations, 50-66 % in most andesitic formations, and from 67-78 % in dacites and rhyolites. Rocks in andesitic formations that have high silica values are the felsic andesites and dacites. Silica variation diagrams in Figures 43 to 45 show the compositional fields for least altered rocks from each formation in each volcanic belt, and Appendix 2D shows plots of individual samples.

The Tumpline Lake subarea shows the strongest bimodal character of all the volcanic belts (Fig. 7). The rocks are dominantly basalts and rhyolites and the amount of andesites is insignificant. A strong bimodal character is not obvious in the Cameron River and Sunset Lake subareas. The Sunset Lake subarea shows the most gradual continuum from basalts through andesites and dacites to rhyolites.

Major oxides

In general, formations within each belt show a progressive increase in the amounts of total alkalis with increasing SiO₂ and decreasing amounts of Fe (total), MgO, CaO, Al₂O₃, MnO, and TiO₂, as would be expected for any transition from mafic to felsic rocks. P₂O₅ does not show any distinct trend with SiO₂. Although Na₂O and K₂O individually show considerable scatter, combined they show a distinct covariance with SiO₂. Individual alkalis have greatest scatter amongst rhyolites, most of which are altered with respect to alkalis. (Note that considerable scatter is removed when only unaltered compositions are used - alkalis are considered to be the major elements affected in alteration of these rocks.) In all belts soda distinctly increases with increasing silica in passing from basalt to andesite and magnesia and lime distinctly decrease.

With the exception of the paucity of andesites in the Tumpline Lake subarea, there is no obvious chemical distinction between rocks of the three volcanic belts: fields of the various rock types in the three volcanic belts overlap with one another. There appears to be a transition of most oxides from basaltic to andesitic compositions.

The Sunset Lake Basalt contains some samples that are distinctly less siliceous than all other basalts. These rocks, which are part of the pale green basalts that lie near the southeast side of Sunset Lake, show the greatest scatter of all oxides and trace elements, suggesting that they are altered.

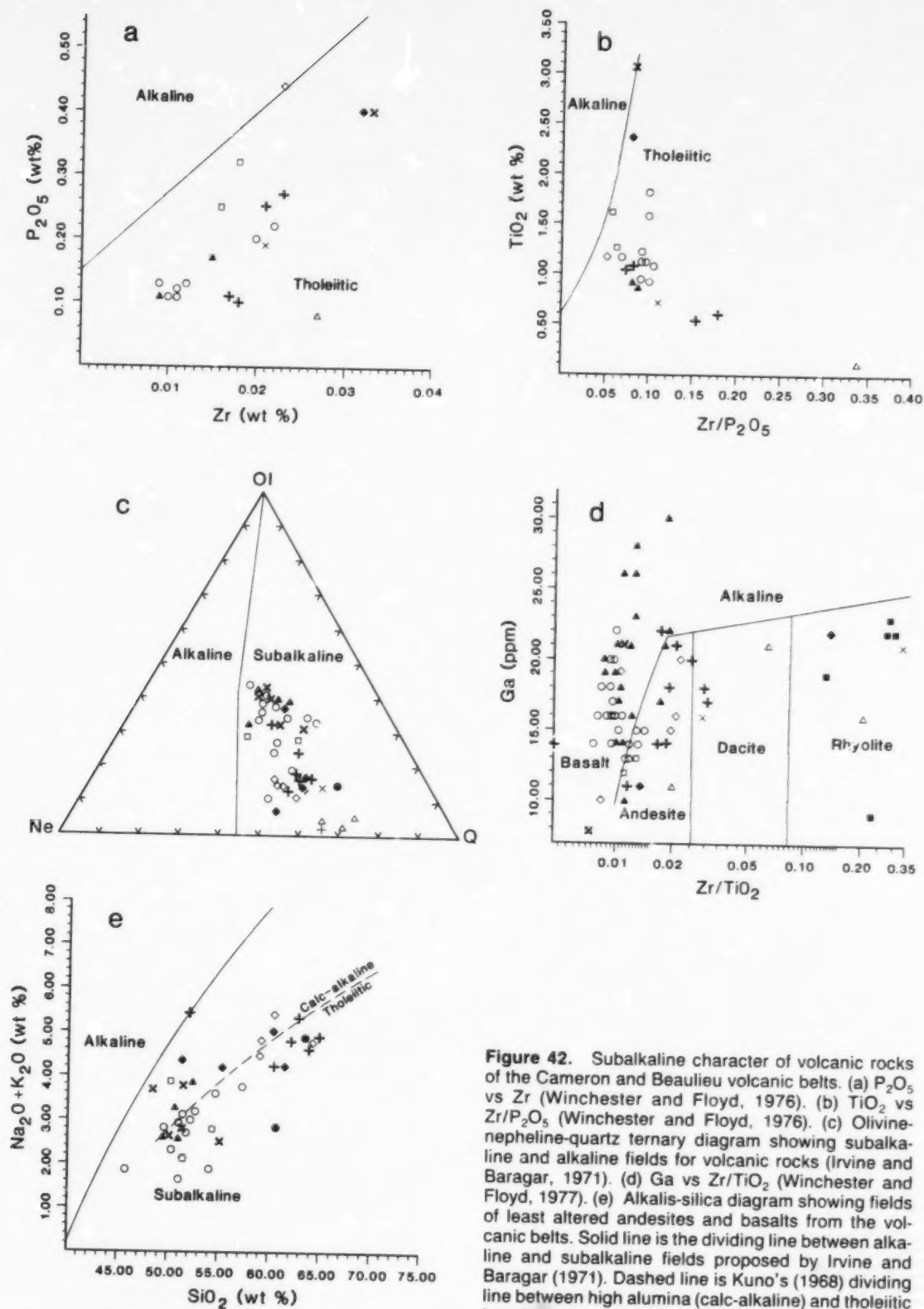


Figure 42. Subalkaline character of volcanic rocks of the Cameron and Beaulieu volcanic belts. (a) P_2O_5 vs Zr (Winchester and Floyd, 1976). (b) TiO_2 vs Zr/P_2O_5 (Winchester and Floyd, 1976). (c) Olivine-nepheline-quartz ternary diagram showing subalkaline and alkaline fields for volcanic rocks (Irvine and Baragar, 1971). (d) Ga vs Zr/TiO_2 (Winchester and Floyd, 1977). (e) Alkalis-silica diagram showing fields of least altered andesites and basalts from the volcanic belts. Solid line is the dividing line between alkaline and subalkaline fields proposed by Irvine and Baragar (1971). Dashed line is Kuno's (1968) dividing line between high alumina (calc-alkaline) and tholeiitic basalts from Japan. Plot symbols indicate formations listed in Appendix 2.

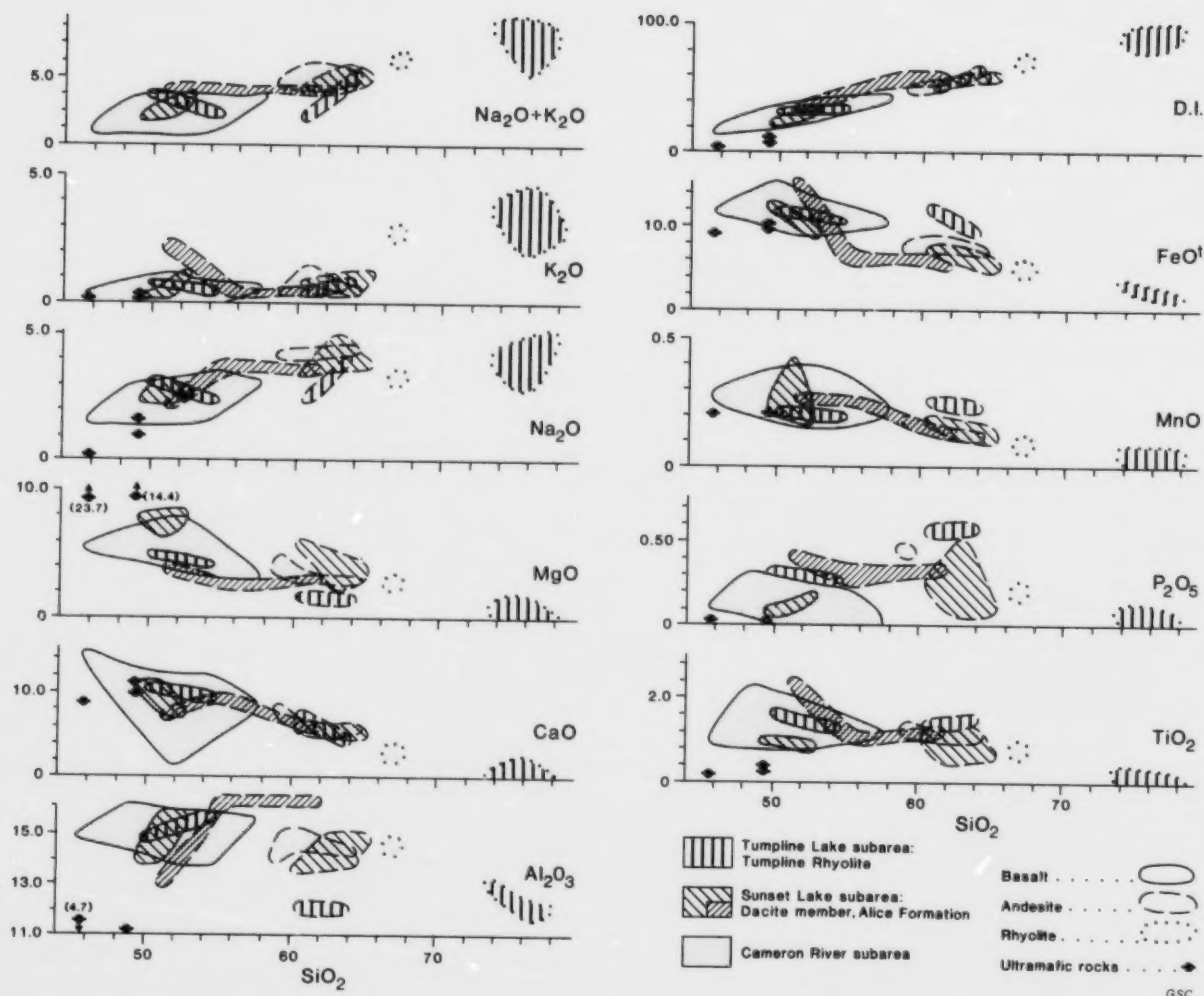


Figure 43. SiO_2 variation diagrams for major oxides (wt. %) in least altered volcanic rocks of the Cameron and Beaulieu volcanic belts.

Very dark green basalts of the Sunset Lake Basalt have lower amounts of Al_2O_3 and MnO but higher CaO , $\text{Fe}(\text{total})$ and MgO than other basalts of the formation. This is reflected in the fact that they have the highest normative colour index of all rocks in the formation.

The andesite member of the Tumpline Basalt appears to be somewhat anomalous in that it is (1) among the most siliceous of andesites, (2) higher in $\text{Fe}(\text{total})$, MnO , P_2O_5 and TiO_2 and (3) lower in Al_2O_3 and MgO than most other andesites. The andesite member plots in a distinct field marginal to or separate from fields of the other andesites (Fig. 40).

Minor elements

Minor elements show a wide scatter for most rock types. In spite of the considerable scatter, Co , V , and Sc tend to decrease with increasing SiO_2 , whereas Pb tends to increase. Fields of distribution for the Turnback Rhyolite, Tumpline

Basalt and andesite member of the Alice Formation must be viewed with caution because the elements Sr , Ba , Cr , Zr , Ni and Zn have not been analyzed for samples 534 to 572 (see Appendix 2B). For this reason, distribution fields for Turnback Rhyolite and the andesite member of the Tumpline Basalt are not shown for these elements in Figures 44. Minor elements do not reveal any distinction between basaltic rocks of the three volcanic belts.

Sunset Lake Basalt shows the greatest scatter in Ga and Ba and contains samples that are anomalously high in these elements compared to other basaltic rocks (these anomalous samples are among the pale green basalts).

Webb Lake Andesite is higher in Sc and Ni and marginally higher in Cr than other andesites. The andesite member of the Tumpline Basalt is higher in Y than other andesites of similar SiO_2 content and lower than most andesites in V . The andesite member of the Alice Formation (Sunset Lake

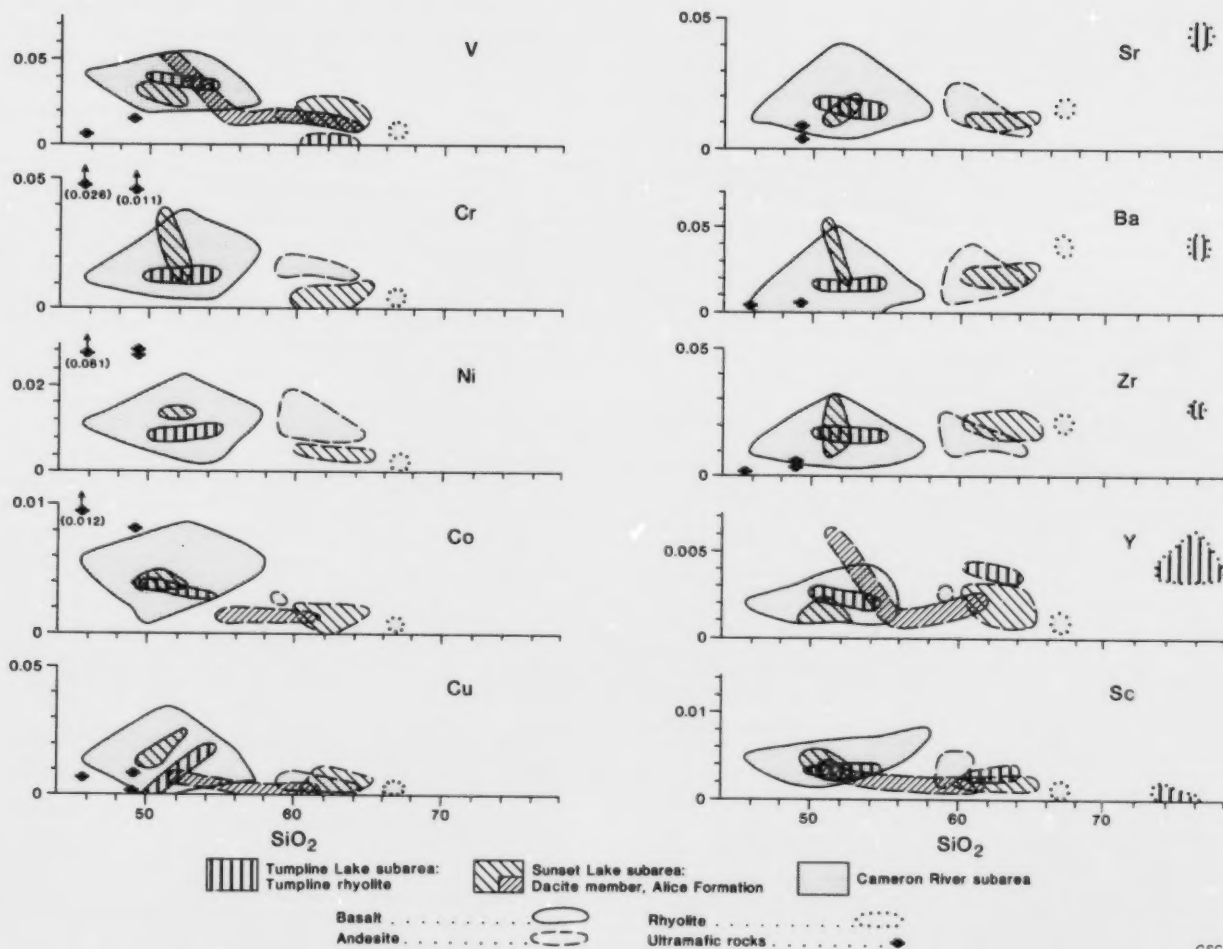


Figure 44. SiO₂ variation diagrams for minor elements (wt. %) in least altered volcanic rocks of the Cameron and Beaulieu volcanic belts.

subarea) shows a greater scatter of values than other andesites for most minor elements and overlap fields of the other andesites. Andesite and dacite members of the Alice Formation overlap in almost every case and are not distinguished in the SiO₂ variation diagrams. The dacite member is not distinctive from the andesite member for most elements except that it tends to be lower (near the limit of detection) than other andesites in Cr and Ni, and higher in Zr.

Rhyolites are difficult to assess because of the paucity of samples. The few data points show a wide scatter in plots of most minor elements. Rhyolites, however, tend to have values lower in Co and Sc than most andesites and basalts, and somewhat higher values in Pb than most basalts. The Sharrie and Turnback rhyolites cannot be distinguished by the available minor element data.

Change of chemistry with stratigraphy

Baragar (1966) carried out the first systematic chemical study of the Yellowknife and Cameron River volcanic belts. He found that the belts showed a small increase in salic components with stratigraphic height, culminating abruptly in acid layers and repeated systematic increases in the iron-magnesium ratios over limited stratigraphic ranges, but no overall systematic variation.

The Cameron River belt was sampled, during the present study, at intervals of 100-400 m to determine stratigraphic changes in chemistry of the lavas with time during development of the volcanic succession. Figure 46 shows the chemistry of one section sampled by Baragar (section 5) and another section about 5 km to the north (section 7-8). Although there does not appear to be any systematic change in individual oxides within the formations, the succession as a whole show general trends from mafic to felsic with increasing strati-

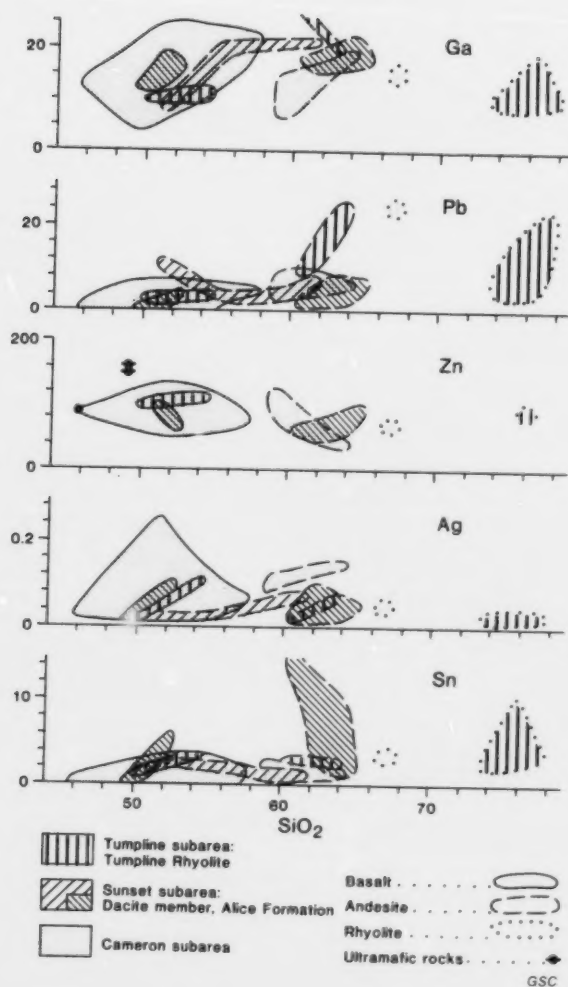


Figure 45. SiO_2 variation diagrams for minor elements (ppm) in least altered volcanic rocks from the Cameron and Beaulieu volcanic belts.

graphic separation from east to west: colour index shows a systematic decline in section 7-8. No systematic change of composition with time, within basaltic and andesitic formations, is revealed by AFM or Jensen plots (Fig. 40).

Figures 47 and 48 show chemical variations between formations in the Tumpline Lake and Sunset Lake subareas. In these diagrams the formations are stacked vertically with respect to relative stratigraphic position in time. The position of samples within each formation does not adhere strictly to the internal stratigraphy of the unit unless indicated otherwise (arrows in Fig. 48). Some samples within a formation are from localities that are widely separated within the volcanic belt and thus no stratigraphic position is known. Nevertheless, Figures 47 and 48 illustrate the change from mafic to felsic with time.

In all three belts the increasing acidity of the volcanics with time is indicated by the differentiation index. In the

Cameron River and Sunset Lake belts and most of the Tumpline Lake belt the enrichment in silica corresponds with apparent increasing stratigraphic height or lateral separation from the earliest products of volcanism. Thus, differentiation trends of each belt are indicated on the SiO_2 variation diagrams, as well as on the AFM and Jensen plots. In the Cameron River and Sunset Lake belts the calc-alkaline trend is indicated by enrichment of alkalis with no corresponding enrichment in iron. A tholeiitic trend is indicated in the Tumpline Lake belt in passing up section from basalts to andesites by increasing iron and titania relative to magnesia with little or no enrichment in alkalis. This is shown in Figures 40a and 40b where the fields for basalts and andesites are plotted.

Discussion

In spite of the deformed and metamorphosed nature of the rocks in the Cameron and Beaulieu belts, their compositions are regarded to be sufficiently similar to modern volcanic rocks to make the following conclusions. (1) All volcanic rocks in these belts are subalkaline. (2) No rocks have alkalic compositions, and ultramafic rocks are rare. (3) The fundamental magma type, which comprises the bulk of the lavas in these volcanic belts, is tholeiitic basalt. (4) Following voluminous subaqueous effusions of tholeiitic magma, the compositions changed along a calc-alkaline trend (leading to alkali enrichment) to form andesites, dacites and rhyolites. (5) Field data suggest strong bimodality of tholeiitic basalts (with minor tholeiitic andesites) and rhyolites in the Tumpline Lake subarea, whereas the Cameron River and Sunset Lake subareas show a transition from tholeiitic basalts through calc-alkaline andesites to rhyolites.

In the Cameron River subarea the change from tholeiitic to calc-alkaline compositions is abrupt at the conformable contact between the Cameron River Basalt and the Webb Lake Andesite. The two formations are chemically distinct and compositions plot in separate clusters in SiO_2 variation diagrams and on opposite sides of dividing lines in AFM plots (Fig. 40a). The Tumpline Lake subarea records an abrupt change from tholeiitic basalts to rhyolites with insignificantly small amounts of tholeiitic andesites.

In the Sunset Lake subarea, the transition between tholeiitic and calc-alkaline compositions appears gradational near the contact between Sunset Lake Basalt and the Alice Formation andesites. In many places, unless sediments define a break within a continuous succession of pillow lavas, the exact contact between the two formations is not precisely defined because of lithological similarities between andesites and basalts. Chemical compositions of the two formations show considerable overlap with each other and cross the boundary between tholeiitic and calc-alkaline fields.

In all subareas basalts are predominantly tholeiites, presumably derived by partial melting and fractionation of the mantle. At some stage of the evolution, magma compositions changed to a calc-alkaline trend and continued to differentiate until the end of the eruptive series (late stage eruption of basalt in the Cameron River and Tumpline Lake subareas, however, represent a second tapping of mantle sources). Such a change may come about by (a) contamination

of tholeiitic basaltic magma by sialic crustal material (as suggested by Baragar (1966) for basalts near Yellowknife and Cameron River), (b) mixing of mafic and felsic magmas, and (c) evolution of a single mafic magma by fractionation under conditions where pressure and water content allowed for production of andesitic magma which continued to fractionate (Osborn, 1959; Green and Ringwood, 1968; Kushiro, 1972; Lambert, 1974)¹. In areas where there seems to be a gradual change of compositions from basalts through andesites to dacites and rhyolites, such as in the Sunset Lake subarea, the two distinct trends do not necessarily indicate two discrete differentiation trends that result from different fractional crystallization processes or different chemistries of primary magmas. Calc-alkaline chemical trends can be derived by mixing of cogenetic magmas whose differentiation trend is initially tholeiitic (Sakuyama, 1983). Chemical features of these magmas are determined by the degree of partial melting of mantle diapirs which in turn depend on the 'stopped depth' of the diapirs (*ibid.*). Magmas segregated from the diapir, rise through the crust, fractionate and extrude to make a composite volcano.

Intermediate calc-alkaline rocks can be formed by hydrothermal alteration of bimodal basalt-rhyolite tholeiites, making them petrographically indistinguishable from metamorphosed andesites and dacites (MacGeehan, 1978). This concept may apply to local parts of the Cameron and Beaulieu belts where (1) andesites are shown to be dominantly altered rocks, (2) the main elements contributing to this alteration are SiO₂ and alkalis, (3) compositions having anomalous minor element distribution are also altered with respect to major oxides, and (4) contacts between tholeiites and calc-alkaline units appear gradational. It is unlikely that this mechanism would apply to produce all the andesites of these belts, because it would require that wholesale hydrothermal metasomatism took place over distances of 25-30 km, not only in areas of felsic centres, but also where such centres are sparse or absent. Also, the metasomatism would have to have been highly selective to produce interfingering of units and sharp conformable contacts. Surely such wholesale metasomatism would be evident in the chemistry.

The Tumpline Lake subarea is anomalous in three ways when compared with the Cameron River and Sunset Lake subareas: (1) felsic rocks that make up 45 % of the belt are almost all rhyolites; (2) south of Tumpline Lake, the oldest volcanic units exposed are the Turnback Rhyolite (the Sunset

Lake Basalt overlies or interfingers with the rhyolites in this area); and (3) andesitic rocks, that make up less than 2 % of the belt, have tholeiitic affinities like the basalts (in Fig. 40 volcanic rocks in the Tumpline Lake subarea show a strong tendency for compositional change along a tholeiitic trend clearly different from the trends in other belts). The highly bimodal character of this belt (abrupt fluctuation from mafic compositions with vanishingly small amounts of andesite) and the apparent tholeiitic differentiation trend among the mafic lavas requires penecontemporaneous eruption from two contrasting magma sources: conceivably tholeiitic basalts and andesites derived from the mantle and rhyolitic magma derived by melting of continental crust. The supracrustal rocks in this subarea are profusely intruded by high level, syndeformational plutons closely related in time to the volcanic rocks. It is possible that rhyolite and plutons were derived from the same source: rhyolitic volcanism being the precursor to Defeat plutons which intruded their own ejecta.

Accepting that granitic plutonism is closely related in time to volcanism, contamination of mafic tholeiitic magma by mixing with granitic magma to form andesitic suites is a likely possibility, particularly in the Sunset Lake subarea where almost all of the high alumina basalts and tholeiitic andesites occur.

Numerous authors have attempted to distinguish magma types that prevail in various tectonic environments in terms of trace element content (Pearce and Cann, 1971, 1973; Cann, 1970; Pearce et al. 1975; Glassley, 1974). Problems that plague the use of such data as a tool to identify paleotectonic environments are: (1) how consistently are magma compositions unique to a particular environment; (2) what is the effect on chemistry of secondary processes such as weathering, hydrothermal alteration, metamorphism and recrystallization; and (3) during Archean times did the same types of magmas prevail in tectonic environments analogous to those of the present. The elements Ti, Zr, Y, and Nb generally remain unaffected by low grade metamorphism (Pearce and Norry, 1979; Cann, 1970; Smith and Smith, 1976). During amphibolitization, Y, Nb and TiO₂ are immobile while P₂O₅ and Zr show a very slight increase (Winchester and Floyd, 1976) suggesting that amphibolites may be classified according to criteria established for fresh basaltic rocks.

Pearce and Cann (1973) pioneered suggestions that ocean-floor basalts, calc-alkaline basalts, low-K tholeiites (island arc basalts) and 'within-plate' basalts can be distinguished chemically. It is less easy to distinguish between continental and oceanic environments where there is considerable overlap (with respect to Ti, Zr, Y, Nb, and P), and no meaningful separation can be made between the two environments (Floyd and Winchester, 1975; Zeck and Morthorst, 1982; Holm, 1982). Furthermore Pearce et al. (1975) discriminated 'primitive' (i.e. those that are not fractionated) subalkaline basalts from ocean-floor environments from those of non oceanic (continental) environments using major oxides on a TiO₂-K₂O-P₂O₅ diagram. They indicated that metamorphosed oceanic basalts will tend towards K₂O enrichment and leave the oceanic field of the diagram, and therefore, if a metamorphosed basalt falls within the oceanic field it is very likely to be of oceanic origin. Some basalts that have erupted in continental settings (i.e. the Tertiary basalts of

¹ Green and Ringwood (1968) suggested from experimental work that andesitic liquid can be derived from basaltic liquid by separation of garnet and pyroxene at pressure equivalents to depth of 100-150 km; dacites and rhyolites, however, can be derived from an andesitic parent only at lower pressures. R. Lambert (1974, p. 434-5) suggested a model for the origin of andesitic suites from Mount Ararat whereby partial fusion of the upper mantle begins in a major lithospheric shear zone. The magma migrates to the top of the fracture zone where, at a lower pressure (ca. 20-25 kb, 20×10^5 - 25×10^5 kPa) and in the presence of a small amount of water, it fractionates (Kushiro, 1972) to produce an andesitic liquid which continues to differentiate along a calc-alkaline trend.

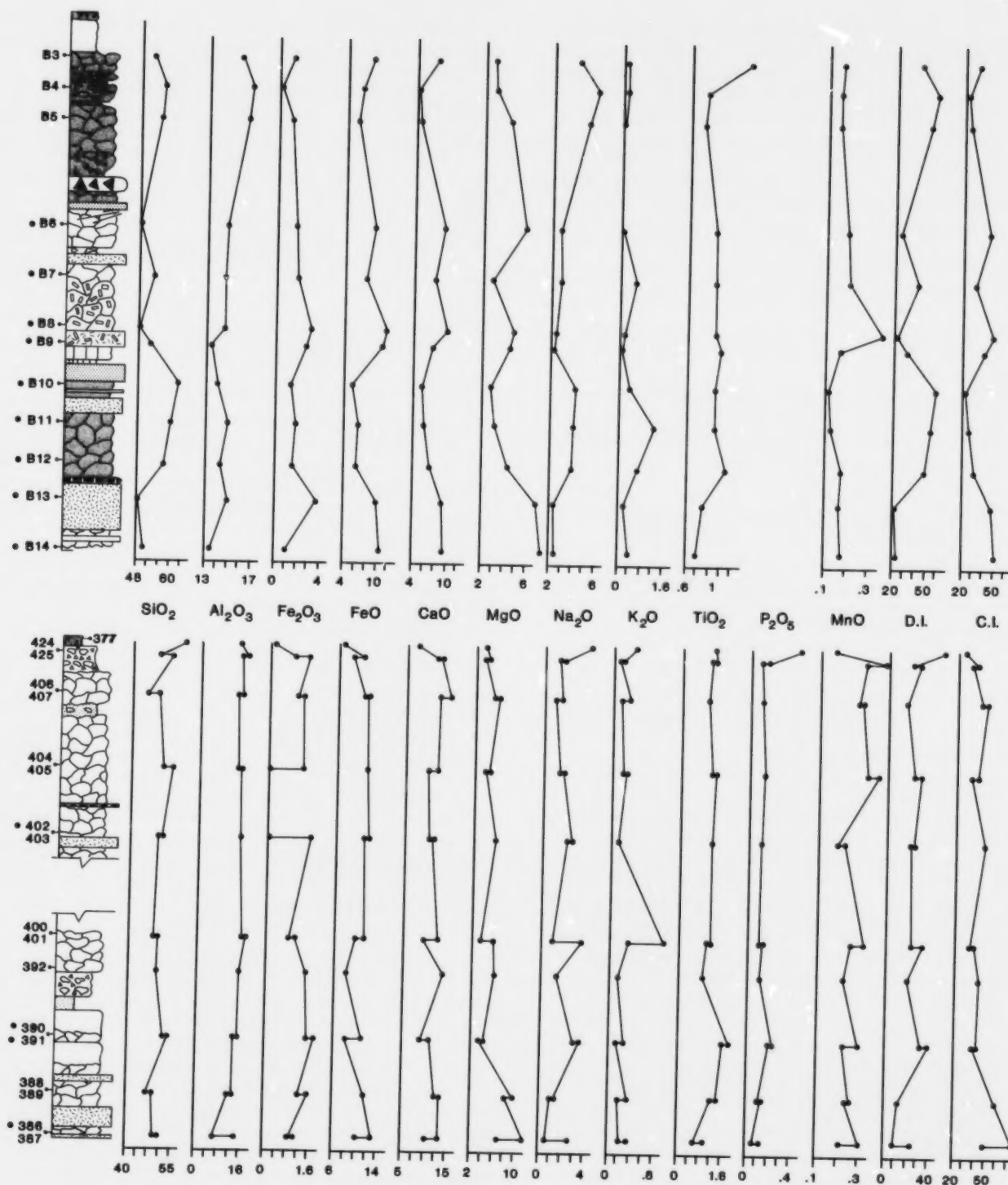


Figure 46. Major oxides and minor elements plotted against columnar sections 5, 7 and 8 of the Cameron River subarea (see Fig. 8, in pocket). Elements followed by * are in ppm: all other elements and oxides are in wt.%. D.I. = differentiation index, C.I. = normative colour index; see Appendix 2c. Symbols in columns as for Figure 8. Grey is Webb Lake Andesite. Dot indicates least altered samples.

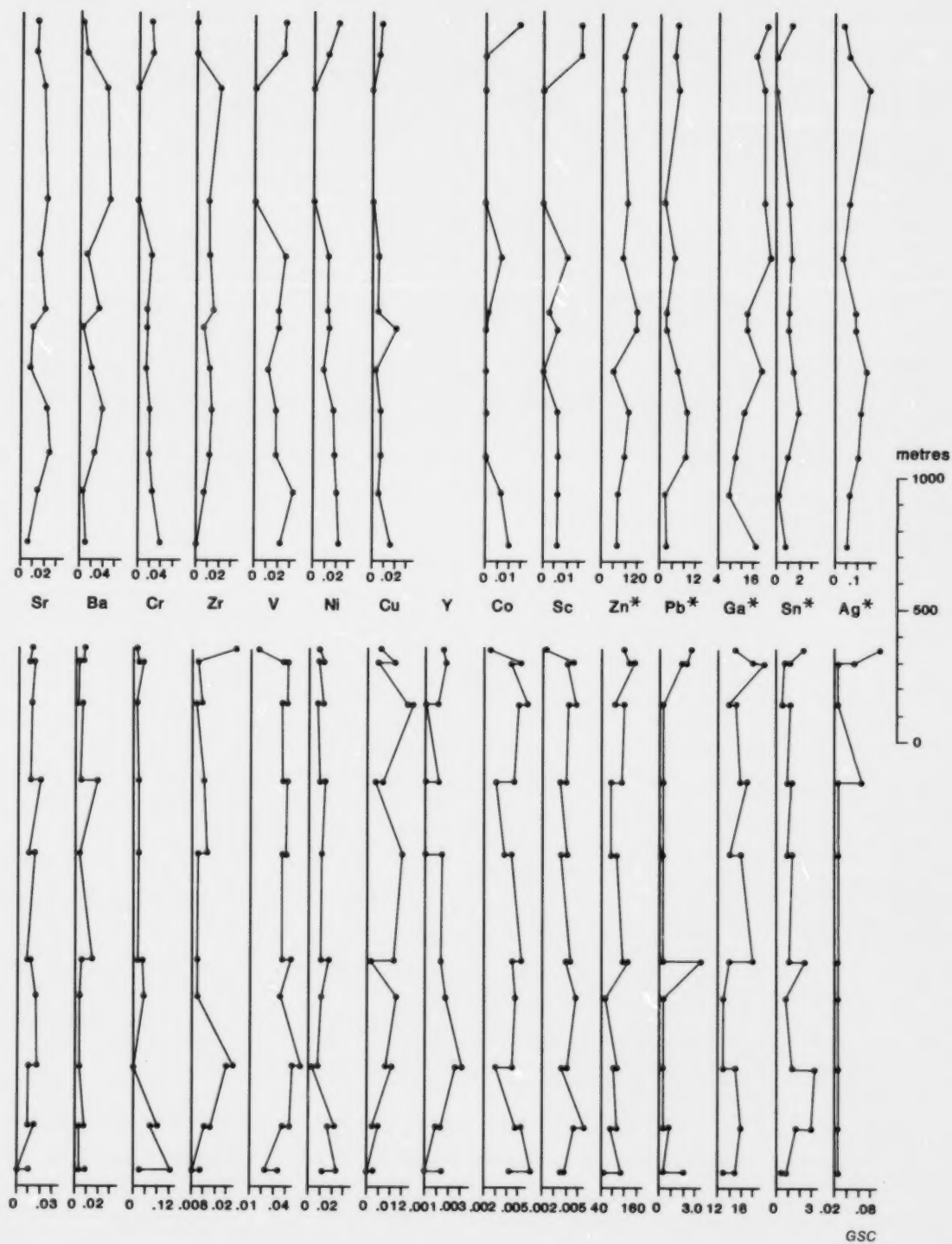


Figure 46 (cont.)

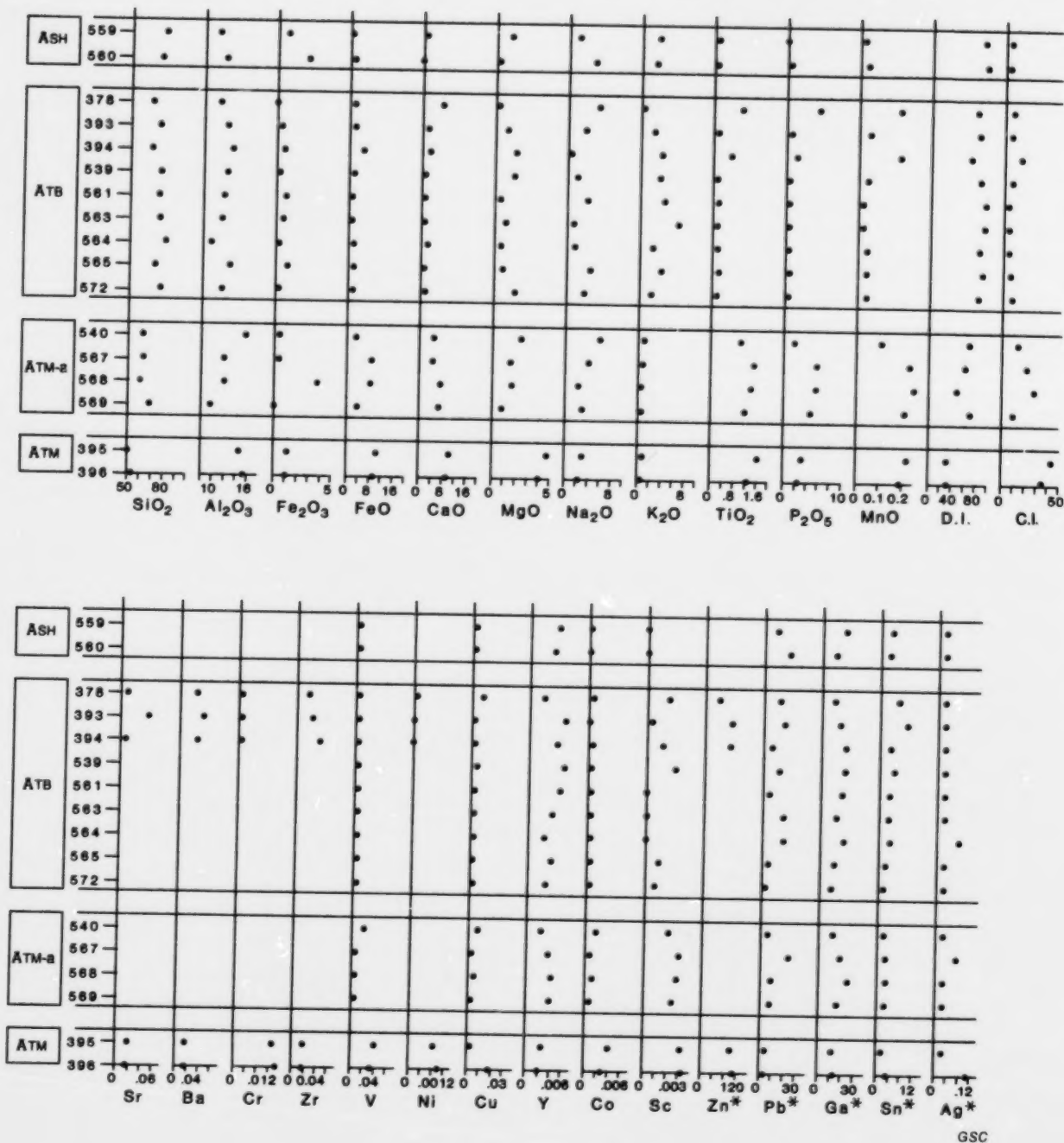


Figure 47. Major oxides and minor elements in volcanic rocks from the Tumpline Lake subarea. Elements followed by * are in ppm; all other elements and oxides are in weight%. Specimen numbers refer to numbers in Appendix 2. Letters and boxes refer to formation unit symbols, as shown in Figures 2 and 3. D.I. = differentiation index, C.I. = normative colour index: see Appendix 2c. Dot indicates least altered samples.

Greenland, tholeiitic lavas of Skye and Mull, and the Deccan traps of India) plot in fields of oceanic basalts on the $\text{TiO}_2\text{-K}_2\text{O-P}_2\text{O}_5$ diagram (Pearce et al., 1975; Winchester and Floyd, 1984). Such basalts may be related to the initial rifting of continents and generation of sea floor (Pearce et al., 1975). As pointed out by Winchester and Floyd (1984) variations in some trace elements may be more likely to reflect magmatic processes than major differences in tectonic setting.

Most basalts from the Cameron and Beaulieu volcanic belts plot within the field of ocean-floor basalts of the $\text{TiO}_2\text{-K}_2\text{O-P}_2\text{O}_5$ diagram (Fig. 49d). The same basalts, when plotted on other discriminant diagrams fall in different fields, depending on which diagram is used (Fig. 49a,b,c.). Basalts from the Cameron River belt, for example, plot mainly in fields of the ocean-floor basalts in a Zr-Ti-Sr diagram, as calc-alkali basalts of island arc setting in a Ti-Zr plot, and are scattered through the fields of within-plate and calc-alkali basalts in a Ti-Zr-Y diagram. The only consistency between the three diagrams is that none of the basalts plot within the fields of low-K tholeiites. Although plots of Pearce and Cann (1971, 1973) seem to be unreliable for determining the tectonic setting of basaltic rocks of the Cameron and Beaulieu volcanic belts, when both major and minor element chemistry are considered, tholeiitic basalts of the Cameron-Beaulieu belts are most similar to those found in ocean floors (MORB), but to a lesser degree also to some island arc and within plate settings. Generally they do not show the enrichment of incompatible elements that is typical of basalts from seamounts. Calc-alkaline basalts are sparse but occur within oceanic tholeiites in the Sunset Lake subarea where calc-alkaline andesites are most abundant and the marked basalt-rhyolite bimodality (as in the Tumpline Lake subarea) is not present. A large part of the volcanic succession in the Sunset Lake subarea has chemistry more similar to that of island arcs than to the other settings of the discriminant diagrams. A clear chemical distinction in all subareas is that no rocks are alkalic and rocks having ultramafic compositions are rare.

STRUCTURAL GEOLOGY

Structural geology presented in this section pertains mainly to structures within the volcanic belts and immediately adjacent metasediments, basement and plutons. Previous workers (Henderson, 1941; Fortier, 1947; Henderson et al., 1972; Henderson, 1985; Fyson, 1975, 1980, 1982) devoted attention almost entirely to folding within the Burwash Formation and recognized that the metasedimentary rocks are complexly folded as a result of more than one and possibly three periods of deformation. Faults and shear zones formed pre-, syn-, and post-volcanism and deformation by folding. Of particular significance are high-strain zones of Archean age marking the contact between the Sleepy Dragon Complex and the volcanic belts. Northerly trending transcurrent faults are probably Proterozoic.

The geology of the Sleepy Dragon Complex is incompletely known and only partial description of its complexity has been attempted (Davidson, 1972) albeit some detailed investigations are in progress along the western margin (Kusky, 1986b, c, d, in press).

This section presents the main structural features for each subarea and the following chapter gives a synopsis of the structural and volcanic history.

Folds

Folds that define prominent patterns in aerial photographs have been described as steeply inclined to overturned isoclinal folds that have gently to steeply plunging fold axes. Although they vary considerably in shape, size and orientation, the most consistent feature is the steepness of axial surfaces variably spaced from several kilometres to less than a few hundred metres (Henderson et al., 1972). Henderson (1985) synthesized and interpreted structural data in metasediments south of Ross Lake over an area of about 750 km² and showed how the northerly trending folds producing complex interference structures with earlier folds swing abruptly to southeasterly trends where they approach and conform to the southwestern edge of the Sleepy Dragon Complex. Fyson (1975) defined three periods of folding in metasedimentary rocks to the west and south of the Cameron River subarea as follows: F_1 , large-scale elliptical culminations and depressions (premetamorphic) defined by opposing facing directions of later folds; F_2 , steeply inclined isoclinal folds with curving axial traces that vary considerably in trend (generally the folds most visible in aerial photographs); and F_3 open to tight upright folds having subvertical axes and a steep axial plane foliation defined by muscovite and biotite that trend northwest to north. He suggested that F_1 folds are consistent with initial development of sinks and intervening upwarps marginal to a diapirically rising mass of granitic basement. The curving trends and overturning of F_2 folds away from basement rocks may have resulted from gravity sliding accompanying this uprise and marginal subsidence. He further considered that regional compression near Cleft Lake deflected folds around a pluton that rose between a second and third phase of deformation (Fyson, 1980).

Volcanic rocks in the Cameron River subarea are not deformed in the same manner as adjacent metasediments. Complex structures, clearly displayed in metasediments, are not evident in adjacent metavolcanics. Recognition of structures, however, in thick pillow lava successions is hampered by the general lack of layered rocks, paucity of marker units, and relatively high degree of metamorphism (amphibolite facies) which tends to mask fine primary features.

Recent detailed structural investigations in the Sunset Lake subarea west of Amacher Lake (Lambert and van Staal, 1987), showed that at least 3 and possibly 4 generations of folds can be distinguished within the Beaulieu River belt. A banded iron-formation, because of its well developed fine layering, provides the best evidence for polyphase deformational history of the volcanic belts. Although mesoscopic folds generally are rare in the enclosing mafic volcanics and dykes, evidence for ductile deformation in the form of foliations and lineations is ubiquitous. Folds are more common in schistose rocks.

Since no systematic structural studies were carried out during the initial mapping of these volcanic belts, fold structures are known only in a generalized way and are highly interpretive in the cross-sections of Figures 2 and 3. The fol-

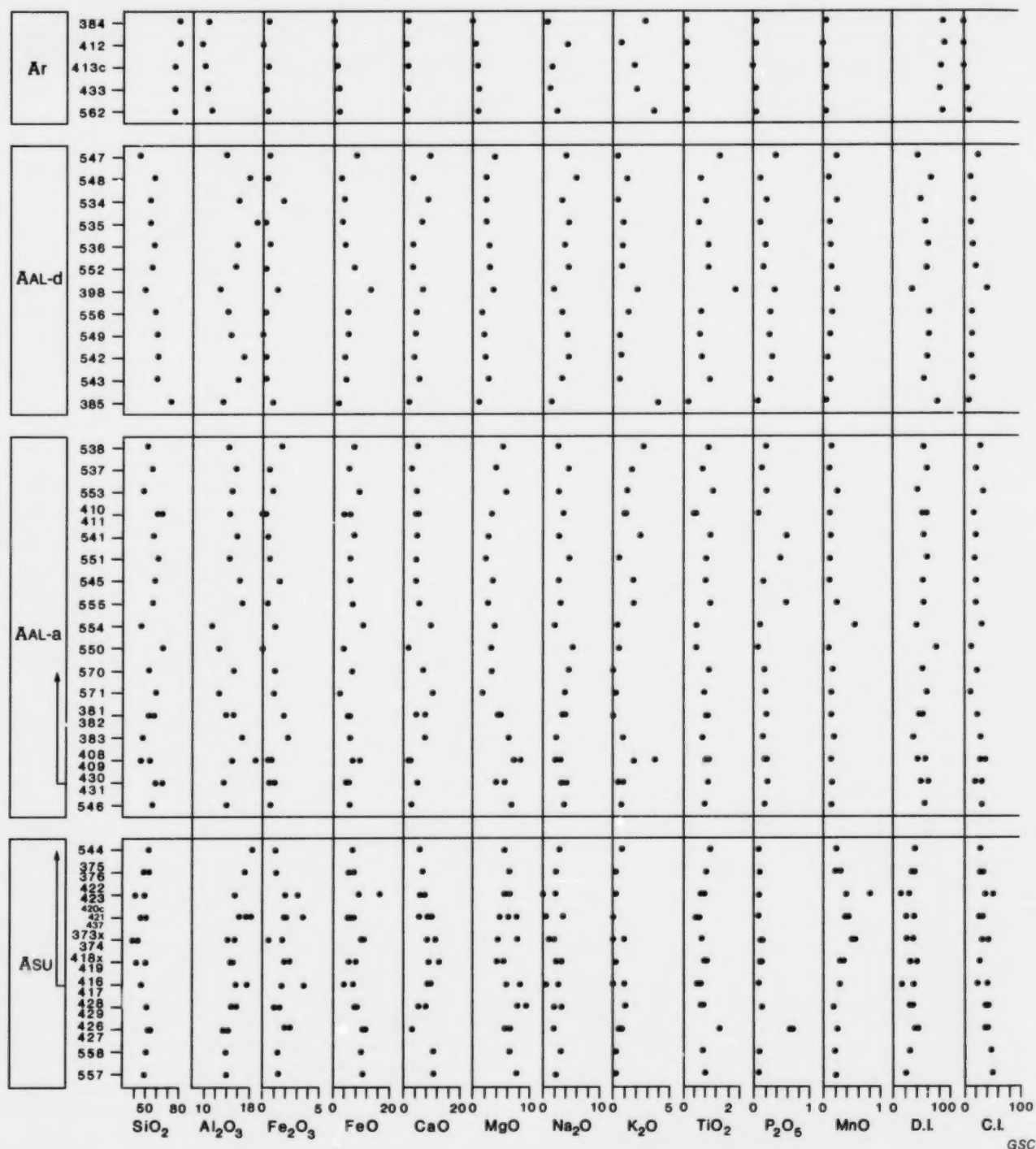


Figure 48. Major oxides and minor elements in volcanic rocks from the Sunset Lake subarea. Elements followed by * are in ppm: all other elements and oxides are in weight%. Specimen numbers refer to chemical analyses shown in Appendix 2 and letters in brackets designate formations as shown in Figures 2 and 3. Arrow in formation boxes indicate direction of younging in a continuous stratigraphic section. Dot indicates least altered samples.

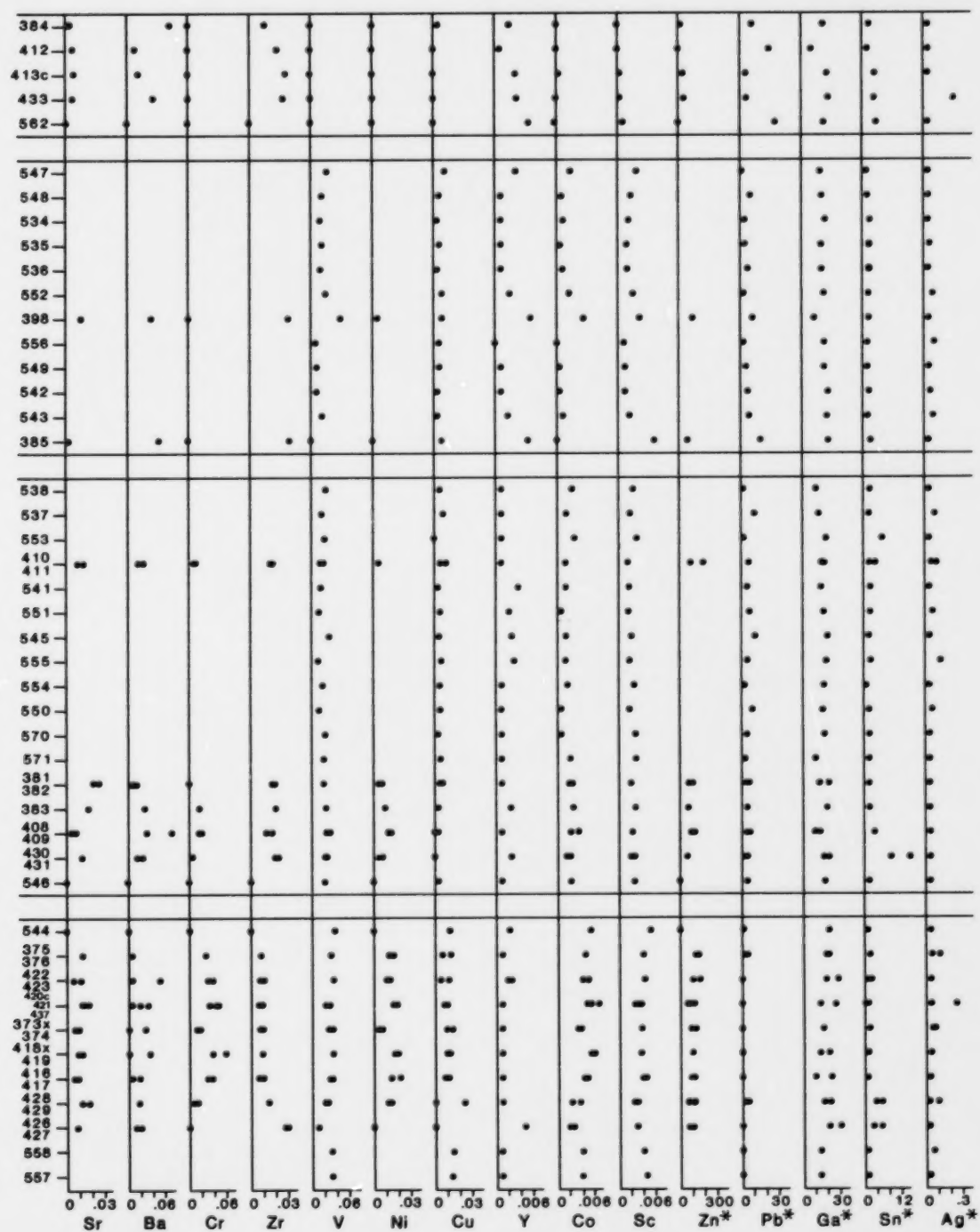


Figure 48 (cont.)

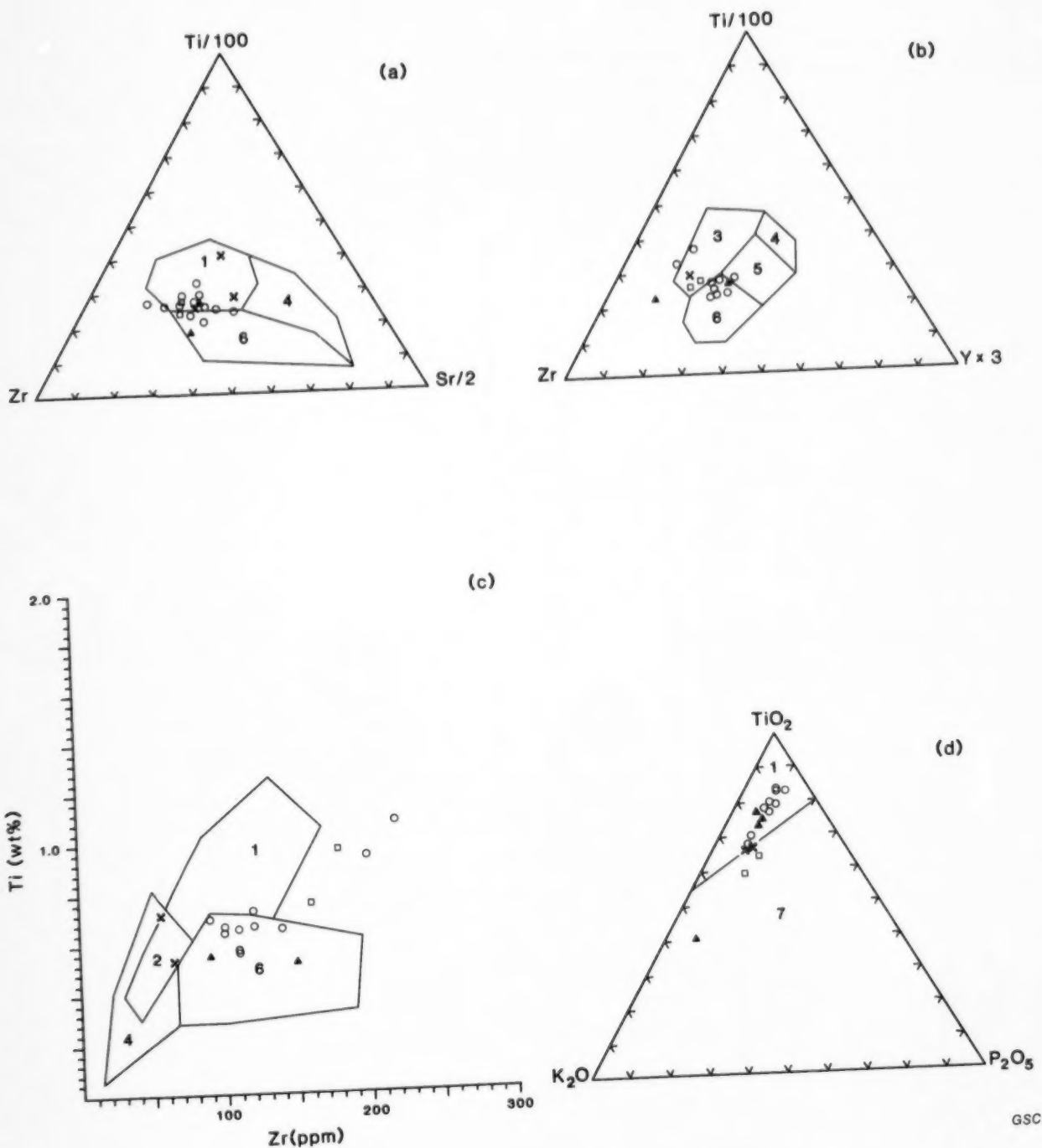


Figure 49. Classification of least altered basalts from the Cameron and Beaulieu volcanic belts according to possible tectonic environment as suggested by Pearce and Cann (1973) for (a), (b), and (c), and Pearce et al. (1975) for (d). Field 1 = ocean floor basalts. Field 2 = ocean floor basalts and island arc calc-alkaline (low-K tholeiite) basalts. Field 3 = within-plate basalts (oceanic island and continental). Field 4 = low-potassium tholeiites from island arcs. Field 5 = ocean floor basalts and island arc volcanics. Field 6 = calc-alkaline basalts of island arcs; Field 7 = nonoceanic basalts.

lowing account describes the main mappable folds and associated minor structures and foliations, loosely referred to herein as early folds, that are equivalent to second-generation folds of Fyson (1975) and Lambert and van Staal (1987). Late folds are small scale northwesterly trending structures equivalent to third generation folds (*ibid.*).

Deformation of the volcanic belts by folding is complicated by the effects of plutons and granitic complexes of regional scale against which the belts abut.

Early folds

The pattern of early folds is different in all three subareas. They range in size from kilometre to centimetre-scale. Shapes range from isoclines to open folds. In the Cameron River subarea the belt appears to be a homoclinal succession in which folds are rarely visible. Folds in the Tumpline Lake subarea are tight, large scale isoclines that trend roughly parallel to margins of granitic terrane to the northwest or wrap around plutons. In the Sunset Lake subarea, map-scale structures appear to be less tightly folded than in the Tumpline Lake subarea, and axial traces trend about parallel to boundaries of granitic terrane to the northeast and northwest of the volcanic belt.

The folds on all scales show thickening of layers in crests and thinning on limbs. The extent of thinning on limbs of large-scale folds is difficult to assess because of abrupt changes in stratigraphic thickness along strike and pinching out of units. Such changes in thickness are particularly common amongst rhyolitic units. Orientations vary from upright to overturned, gently to steeply plunging isoclines and open folds. Axial traces have variable orientations but generally conform to boundaries of adjacent plutons or regional granitic complexes.

Main foliations (defined by a cleavage or schistosity in rhyolites) and general direction of flattening of pillow lavas, generally parallel axial planes and layering on limbs of isoclines. Commonly, the degree of flattening of pillow lavas changes towards the closure of folds. In the large isocline at the southeast side of Tumpline Lake, for example, pillows on horizontal surface vary from intensely flattened on fold limbs to equant forms in the nose, which have been stretched into rods that plunge about 80° in an easterly direction.

Degree of flattening is judged from the apparent ratios of length to width of pillows on horizontal surfaces. These ratios are only rough indicators of the flattening strain in most places, because the primary shapes, dimensions and orientations of pillows are generally unknown. Furthermore, pillows in many areas are elongated (stretched) in vertical sections. The most intensive flattening observed occurs along the limbs of isoclines within the anticlinorium south of Detour Lake. In this area the pillows range from weakly deformed to flattened ribbons 2-10 cm thick and several metres long. Ratios of length to width range from 12:1 to 50:1 (Fig. 14).

In most tightly folded areas, highly elongated pillows and breccia fragments generally define a lineation that plunges steeply within the axial plane. In at least some places this lineation is perpendicular to shallow plunging fold axes. Elongation or "stretching" of pillows may be related to early

deformation that formed the high strain zones near volcanic-granitic boundaries (Lambert and van Staal, 1987) and some of the discrete shear zones within volcanic belts.

Lineations in massive amphibolite dykes and sills are defined by preferred orientation of hornblende crystals and quartz-feldspar ribbons. They generally parallel contacts of dykes and at least in some places are coaxial with the elongated pillows in adjacent basalts.

Small-scale early folds are tight to isoclinal, nearly similar folds with amplitudes commonly 5 to 10 cm and less. They commonly define "S" and "Z" asymmetrical structures in the laminations and eutaxitic foliation of rhyolitic units, and digitations in thin rims of pillows and in layering within pillows on the limbs of larger folds. In some localities, tight minor folds in the foliation, thin felsic dykes and veins within large massive amphibolite units (lava, sills or dykes) indicate significant deformation, whereas the units otherwise appear superficially undeformed.

Late folds

Late folds deform planar structures related to early folding. In all subareas their axial traces and cleavage trend north to northwesterly, and in most places cut across axial planes of early folds at moderate to high angles. Late folds are small-scale structures that range in amplitude from a few centimetres to rarely more than a few metres, and vary from gentle undulations or open concentric folds to tight crenulations of layering and foliations defined by the early fold axial plane schistosity or flattened pillows, pillow rims and clasts. Locally, hook folds (Type 3 interference pattern of Ramsay, 1967) formed where late folds have deformed small-scale isoclines. The foliation that defines axial planes of late folds varies from a weak discontinuous fracture to a closely spaced penetrative crenulation cleavage or schistosity.

Axes of late folds generally plunge steeply. At one locality near Webb Lake the intersection of axial plane cleavages related to both early and late folds divides the outcrop into pencil-like slivers parallel to the axes of late folds.

The trend of late folding was not markedly influenced by earlier major deformation or emplacement of plutons. The constant trends in all subareas imply a regional shortening in a west-southwesterly to east-northeasterly direction (Henderson, 1985).

Folds in the Tumpline Lake subarea

Three dominant patterns of early folds in the Tumpline Lake subarea include a northwest-trending isoclinal anticlinorium south of Detour Lake, north-northeast-trending isoclines between Tumpline and Turnback lakes, and deflection of folds around plutons south and northeast of Tumpline Lake.

The isoclinal anticlinorium south of Detour Lake is about 3 km wide and passes northeastward into an antiform-synform pair. Wavelengths range from 1.5 to 2.5 km and axial traces parallel the boundary of the granitic complex to the north. The axial traces generally trend northwesterly (ca. 300°) but curve slightly northward near the nose of the anticlinorium and southward near the southeast end. The folds are trun-

cated eastward against a northerly trending, crosscutting pluton of Prosperous Granite. The anticlinorium is a slightly twisted gently northwesterly plunging structure. The steeply inclined axial surfaces change dip along the length of the anticlinorium so that the folds are overturned toward the southwest near the nose, and toward the northeast near Devore Lake (see cross-section, Fig. 2).

Plunges of these folds are not well constrained and the three-dimensional geometry has not been clearly defined. In Figure 2, a schematic profile (section line S-T-U-V-W) along the axial surface of the anticlinorium shows the stratigraphic relations in this part of the area but does not attempt to portray the structural complexity. Figure 56 shows the complex fold interference pattern of the polydeformed volcanic belt in the southern part of the Tumpline Lake subarea.

Between Tumpline and Turnback lakes early folds are tight isoclines, which are slightly overturned to the northwest and plunge toward the northeast. The axial traces trend north-northeasterly parallel to the boundary of the Sleepy Dragon Complex to the west.

South of Tumpline Lake axial traces of major folds curve to the southeast, in rough conformity with outlines of granitic plutons to the east. All volcanic units in this area wrap tightly around a large pluton (about 4 km across) and two smaller plutons (ca. 0.5 km across) to the north. The few stretching lineations recorded in volcanic rocks plunge steeply away from the plutons.

Within 2.5 km of the Sleepy Dragon Complex (between Tumpline and Turnback lakes), folds are regional scale, high amplitude isoclines that contain substantial thicknesses of rhyolite in the hinges. Between 2.5 and 4 km from the granitic complex, the folds diminish eastward into a series of smaller amplitudes and wavelengths upright folds (section J-K, Fig. 2). Axial traces of these folds (marked by the trend of rhyolite units where the wings of the belt merge) curve to conform to the boundary of the pluton to the east. At this point the major northeasterly fold structure changes abruptly to a westerly plunging antiform cored by a body of the Defeat Plutonic Suite.

Near Rex Lake the outcrop pattern of the volcanic rocks describes a large northeasterly plunging anticline where they wrap around the northern end of an elongate pluton of Redout granodiorite.

This fold pattern is influenced by the boundaries of the Sleepy Dragon Complex and plutons of the Defeat Plutonic Suite which intruded the volcanic belt either before or during deformation. It is suggested that the conformity of layering to these bodies developed during early emplacement of plutons (see Defeat Plutonic Suite) and that subsequently folds were tightened and moulded around the plutons and flattened against the rigid basement.

Folds in the Sunset Lake subarea

The valley of Sunset Lake and the Beaulieu River to the north marks the axis of a major synclinorium down the centre of the volcanic belt in this subarea.

Folds within the Sunset Lake Basalt are subject to interpretation because of the general lack of good marker horizons and reliable top determinations. In places, however, local clastic units indicate bedding tops and massive amphibolites (flows or sills) define fold closures.

South of Sunset Lake, the northerly trend of units within the belt swings abruptly to the southwest. The structure near the south end of Sunset Lake is complicated by lenses or domes of rhyolite and basalt. Distortion of the major synclinal structure is probably related to a granitic pluton to the south.

The central synclinorium passes westward into an anticlinorium that envelopes the large elongate pluton of Amacher Granite. Folds within the anticlinorium near the north end of the pluton plunge gently northward, in contrast to the general southerly plunge of the synclinorium to the east. Around the southern end of this pluton the disposition of major volcanic units and the contact between the Burwash sediments and the volcanic belt define the southerly plunging anticline. Near Amacher Lake, the volcanic belt is interpreted to be deformed into a series of north-northeasterly trending folds dominated by a syncline. Near Rex Lake, an elongate pluton of Redout Granite cores a major anticline that plunges north-northeasterly.

The portion of the belt south of Payne Lake is interpreted as a southerly plunging anticlinorium that has segments displaced by northeast-trending faults. Folds in this area (Fig. 3) are highly interpretive, based on 1) map pattern, defined by alternating felsic (or sedimentary) and mafic units that suggest fold repetition, 2) extrapolation of structures from the north, and 3) attitudes of schistose foliations, assumed to represent axial plane foliations. This portion of the belt comprises a series of upright to overturned folds whose axial traces parallel the general southeasterly trend of the belt. Southward pinching out of mafic volcanics and northward pinching out of felsic volcanic units is consistent with a pattern expected in a folded succession where stratigraphically lower rocks (mafic volcanics) are preserved in the crests of anticlines and the overlying felsic units and sediments are preserved in synclines. The abrupt change in width of the belt as well as the marked increase in the proportion of sediments to volcanics within the belt, across the northeasterly trending fault south of Payne Lake, is consistent with the higher stratigraphic level at the crest of the anticlinorium that is downdropped on the south side of this fault. Furthermore, the sediments in the volcanic belt south of the fault are more similar to Burwash sediments than to sediments in the belt north of the fault. This would be expected of sediments higher in the stratigraphic succession, where volcanic influences are diminishing as volcanism wanes.

The north-northeast-trend of folds in this subarea is interpreted to result from deformation of the volcanic belt against the Sleepy Dragon basement terrane and interaction with the syndeformational Amacher pluton. It is proposed that during folding, rocks progressively deformed around the pluton which controlled the position of the anticlinorium.

West of Amacher Lake (Fig. 22) Lambert and van Staal (1987) defined 3 generations of folds (F_1 , F_2 and F_3) from overprinting relationships mainly in iron-formation within

the Sunset Lake Basalt. First generation folds (F_1) generally are asymmetrical isoclines that fold millimetre- to centimetre-scale compositional layering of the iron-formation. Although F_1 fold closures are rare in the basalts, evidence for this deformation is indicated by a phyllonitic foliation (S_1). Locally F_1 folds are thinned and boudinaged along their long limbs and are rootless. Fold plunges vary from steep to moderately shallow. Some F_1 folds are markedly noncylindrical and closely resemble sheath folds (Lambert and van Staal, 1987, Fig. 7c).

Second generation folds (F_2) are tight to isoclinal, generally asymmetrical, and upright. Close to contact with the Sleepy Dragon Complex, however, F_2 folds are locally overturned toward the west. A crenulation cleavage (S_2) axial planar to F_2 folds is present in F_2 -folded biotite schists and serpentinite. West of Amacher Lake, F_2 folds plunge either northerly or, more commonly, southerly. Locally, they are moderately (10–50° degrees) doubly plunging structures. The tight or isoclinal, intrafolial character of F_2 folds causes transposition of all earlier foliations into parallelism with F_2 axial planes. The regional foliation is therefore an S_2 transposition foliation. Overprinting between F_1 and F_2 folds predominantly gives patterns diagnostic of Type 3 (coaxial refolding) interference (Ramsay, 1967, p. 53), however; mushroom type (Type 2 of Ramsay) interference also occur (Lambert and van Staal, 1987, Fig. 8E). Third generation folds are open to tight, asymmetrical, generally with north-northwesterly trending axial planes. In strongly foliated rocks, such as the serpentinite, they are generally kinks or chevron folds with sharp angular hinges. Elsewhere, in more competent rocks such as amphibolites, the folds can have rounded hinges. F_3 folds have an axial plane foliation which varies from a crenulation cleavage that fans around the hinge to a kinkband boundary like fracture cleavage. F_3 fold axes have variable plunge. In this area F_3 folds dominantly have Z-asymmetry.

Folds and deformation in the Cameron River subarea

The Cameron River volcanic belt does not show an obvious macroscopic fold pattern within itself or with adjacent sediments as displayed in the other subareas. The complex pattern of folds in the adjacent Burwash sediments (as shown by Henderson, 1985, and Henderson, 1941) does not appear to continue into the volcanic belt. Instead the folds swing to conform to the trend of the volcanic belt.

The belt superficially appears to represent a homoclinal succession against the granitic terrane to the east and is steeply dipping and overturned to the west: with few exceptions, pillows face westerly. Marker horizons are few within this extensive succession of pillow lavas. The general lack of repetition of units, consistent facing directions across the belt, and lack of Burwash sediments folded into the volcanics, suggest that the belt does not contain major folds. Some internal folding is suggested by: (1) reversals in facing directions or attitudes of flattening in pillow lavas, particularly along the eastern side of the belt (east of Fenton Lake and near Webb Lake); (2) multiplicity of rhyolite units (east of Fenton Lake); and (3) narrow dykes that converge to a wide unit which could be the noses of gently plunging folds.

Penetrative strain persists throughout the belt and almost all flattening, high strain zones and stretching lineations are steep and have the same general trend as the volcanic belt as a whole or as the eastern boundary of the belt.

Although initially pillows probably had a tubular form, the following features indicate that elongation and preferred orientation are mainly, the result of tectonic deformation (i.e. represent a stretching lineation): (1) direction of elongation of all features is uniform in any particular outcrop; (2) not only pillows show pronounced elongation, but also fragments in pillow breccias, felsic clasts incorporated within pillows, and clasts in conglomerate within the pillowed succession.

In parts of the Cameron River belt, pillows and fragments show a prominent elongation even though they are not apparently distorted perpendicular to elongation. In these areas, pillows and breccia fragments are equant on horizontal surfaces, but highly elongate in vertical section. Ratio of length to width are in the order of 4:1 to 6:1. Elongation of pillows and clasts generally plunges steeply (65–80°) easterly to southeasterly, except in areas of local folding where the plunge is westerly or variable. In areas of intense flattening of pillow lavas the direction of elongation and plane of flattening appear to lie within foliation defined by cleavage and schistosity.

In the Cameron River subarea evidence of internal folding is minimal. Extreme flattening and elongation of pillows, however, are evidence that the belt is thoroughly deformed. Fine crenulation of pillow rims within the planes of foliation, protomylonitic textures on a microscopic scale in rhyolite, and a strong foliation in almost all rocks are consistent with penetrative strain. In places along the eastern margin of the volcanic belt, strain is so intense that pillow structures, massive lavas and dykes are undistinguishable. Steep axial planes of isoclines in adjacent Burwash sediments, as well as steep foliations throughout the volcanic belt, suggest that main shortening was horizontal. The pattern of strain along the eastern side of the belt clearly relates to the zone of high strain along the complex boundary between the Cameron River volcanic belt and the Sleepy Dragon Complex (see Sleepy Dragon shear zone).

The present steep plunge of stretching lineations ubiquitous in the volcanic belt could have had low to moderate attitudes originally, but were subsequently rotated to steep orientations.

Structures within the belt could also be accounted for by processes of pure and simple shear of pillow lavas against a buttress, as illustrated by the models in Figure 50.

In view of the multiple periods of deformation defined in the Sunset Lake subarea, and the evidence for abundant internal deformation and at least some folding in the Cameron River belt, it is likely that detailed structural studies will also reveal multiple folding in this subarea.

Archean faults and shear zones

Faults and shear zones (Fig. 51) play a critical role in understanding the tectonic history of the Cameron-Beaulieu volcanic belts. Structures of Archean age include prevolcanic

faults in the Sleepy Dragon Complex, synvolcanic faults, major syndeformational shear zones along boundaries between volcanic belts and the Sleepy Dragon Complex, smaller scale shear zones within the volcanic belts and at the boundary between volcanic and overlying sediments which

may be related both to marginal shear zones and to folding. Post deformational transcurrent faults are probably Proterozoic.

Prevolcanic faults

The only faults that could be interpreted as prevolcanic are the north to northwesterly trending faults that displace the western side of the Sleepy Dragon Complex and account for its irregular blocky outline. Two of these faults near Webb and Patterson lakes displace the granitic boundary by 1500-1800 m but do not continue through the volcanic belt. The western boundary of the volcanic belt does not reflect the irregularity of the eastern side. The few of those faults that do penetrate the volcanic belt show much less movement than that indicated by the displaced granitic boundary. The more northerly trending surfaces of this irregular boundary also may have been a faulted surface or even an erosional unconformity, but ubiquitous high strain along this boundary masks its previous nature.

The irregular, angular boundary is interpreted as a relic of an original surface against which the volcanic belt formed or deformed. The slight penetration of faults in the Sleepy Dragon Complex into the volcanic belt suggests reactivation of these surfaces possibly during volcanism, but more likely during major deformation of the belt. If, however, the volcanic belts are allochthonous, then these faults are pre-emplacement and not necessarily pre-volcanic.

Synvolcanic faults

Faults within the volcanic pile interpreted as synvolcanic, offset volcanic formations, but do not continue into the basement or Burwash sediments. In the Cameron River subarea westerly trending faults cause prominent topographic lineaments and show maximum displacement in the order of 250 m, but mostly have displacements of less than 50 m. Many do not extend across the entire width of the belt. West of Webb Lake, for example, they displace units within the eastern and central parts of the belt but do not displace units near the western side. Some of the most extensive westerly faults have slightly curved trends. Nearly vertical shear zones within, and roughly parallel to, the volcanic belt are ubiquitous. At least some of these may mark faults active during volcanism.

The present low relief and abundant postvolcanic strain prevents recognition of primary fault scarps. Evidence that fault scarps existed at the time of volcanism is circumstantial. Units of rhyolite and rhyolite-basalt breccia are interpreted as debris deposited in fault depressions in the underlying pillow lava succession. Reactivation of these faults during subsequent deformation produced the present sheared contacts. Similarly, numerous thick lenses of basaltic pillow breccias are interpreted as talus aprons derived from topographic highs, either constructional volcanic scarps or tectonic scarps. A conceivable source for coarse granitic or mixed volcanic-granitic conglomerate of the Raquette Lake Formation is from fault scarps that exposed granitic basement where the volcanic belt tapers out. Similarly, volcanic-granitic conglomerate within the Sunset Lake Basalt could

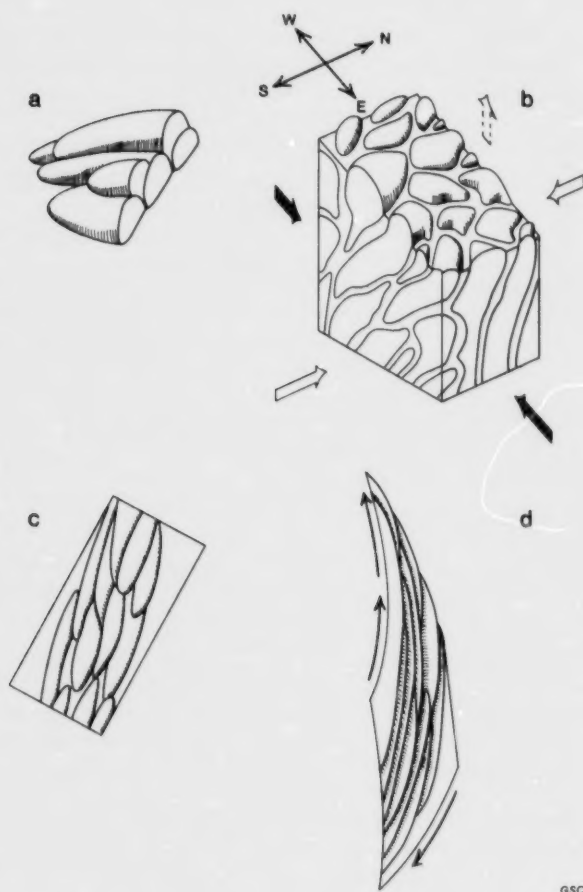


Figure 50. Models of deformation of pillow lavas by processes of pure and simple shear in the Cameron River belt. (a) Before deformation pillows formed a pile of gently plunging tubes. (b) Deformation of pillows in (a) by pure shear (sketch from plasticine models): solid arrows — direction of maximum compression; open arrows — stress imposed by confinement of boundaries (length) of belt in northerly direction; dashed arrow — direction of maximum extension. Pillows are stretched, steeply plunging bodies. On horizontal or shallowly dipping surfaces they have almost equant forms that face in the same general direction. Hatched pattern indicates original bottoms of pillows. (c) Mass of steeply plunging pillows after deformation by pure shear. (d) Deformation of (c) by simple dextral shear along steep upward curving surfaces produced highly elongate forms with regular steep plunge and slightly overturned. Extreme deformation tends to produce pseudoconformable relationships, so that layers and dykes within the pillowed succession become almost parallel to the trend of elongation and flattening.

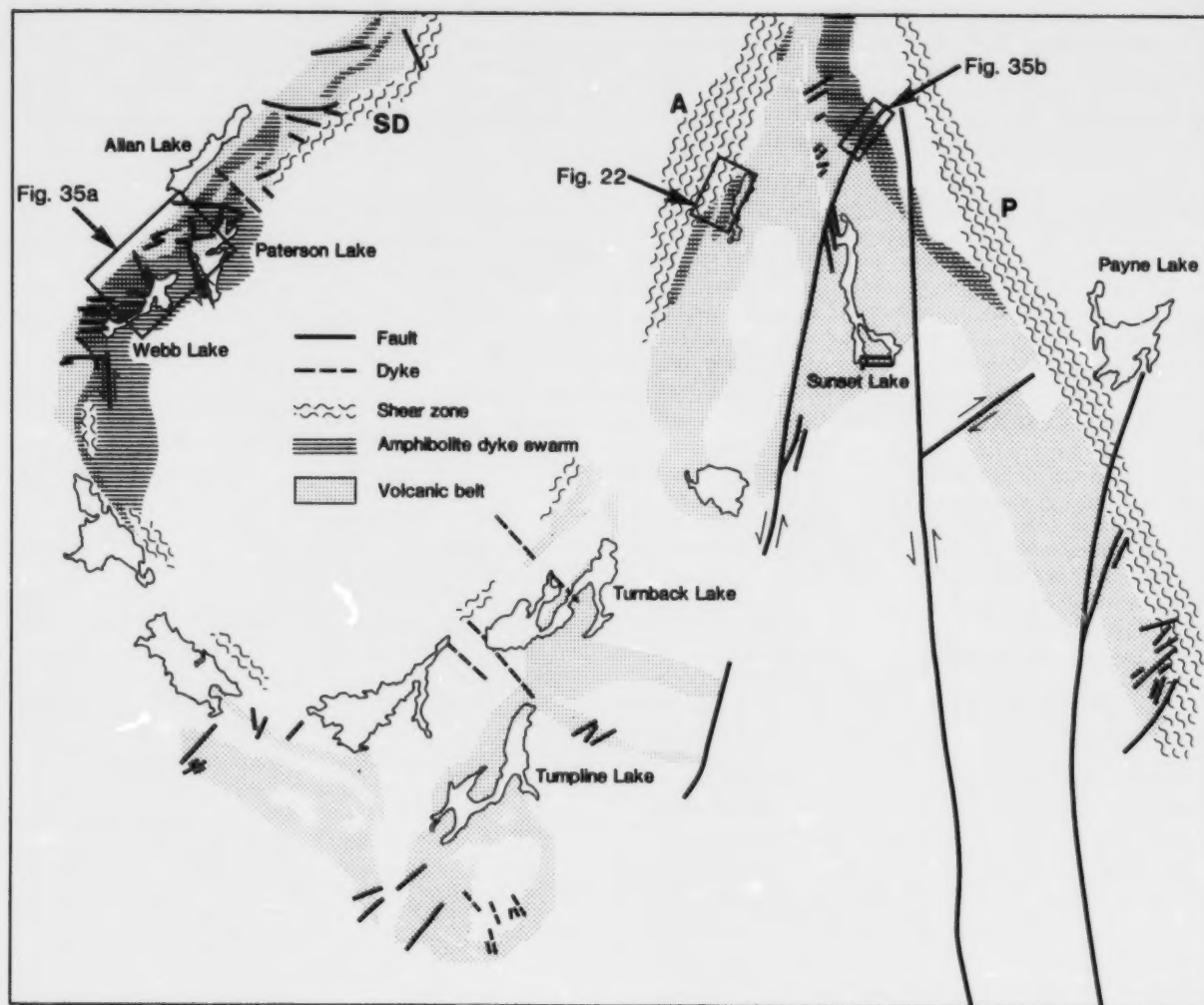


Figure 51. Faults, shear zones and major dyke swarms in the Cameron - Beaulieu volcanic belts. SD = Sleepy Dragon shear zone, A = Amacher shear zone, P = Payne shear zone.

be derived from synvolcanic faults that exposed granitic terrane.

Indeed, it would be highly unusual that such voluminous effusions of lava, erupting through broad zones of fracturing with resulting rapid accumulation of volcanic piles, would take place without concomitant fault displacement.

Although granitic highs may be considered sources of granitoid blocks in some volcanic breccias in the northern parts of the Cameron River belt and in parts of the Sunset Lake subarea, their occurrence within shear zones permits derivation by tectonic processes.

Synvolcanic faults may be related to subsidence during growth of the volcanic pile. Some may be extensions of faults in the underlying substratum. Presumably synvolcanic faults were reactivated, and possibly extended farther than their original dimensions through the volcanic succession, during folding and flattening of the belts. Some would become curved during the penetrative strain that effected the entire belt.

Shear zones and syndeformational faults

Postvolcanic faults and shear zones interpreted as coeval with, or post-dating, folding include major high strain zones along boundaries between the Sleepy Dragon Complex and volcanic belts, discrete shear zones within the volcanic belts, and narrow- to broad-zones of cataclasis and/or ductile strain.

A broad zone of shearing within the volcanic belt along the valley of Sunset Lake and the Beaulieu River to the north, is interpreted to represent movement during folding along the axial zone of a major synclinorium. In this zone, textures of the volcanics indicate both brittle (microbreccias) and ductile deformation. The net effect is protomylonitization (Fig. 28) of felsic lavas and domes which caused outcrops to have megascopic textures resembling pyroclastic rocks (see Alice Formation). The foliation, in part defined by lenticulation in protomylonites, is steeply dipping and northerly trending along the projected axial surface of the synclinal structure whose core lies along this valley. Northerly trending strike-

slip faults near the north end of Sunset Lake account, however, for some of the strain observed in this movement zone.

Similarly, narrow mylonite zones in highly deformed pillow lavas (andesite member of the Tumpline Basalt, Fig. 14a) occur in the core of the anticlinorium south of Detour Lake.

The Sleepy Dragon Complex is bounded by extensive zones of mylonite that involve both granitoid gneisses and the volcanic belts (Fig. 51). The zones are referred to here as the Sleepy Dragon shear zone in the Cameron River subarea and the Amacher shear zone along the eastern side of the Sleepy Dragon Complex, in the Sunset Lake and Tumpline Lake subareas. Whereas these two zones may be coeval in part with some of the folding, the Payne shear zone, near the eastern side of the Beaulieu River volcanic belt, post-dates the latest folding and emplacement of the Meander Lake Plutonic Suite.

Sleepy Dragon shear zone

This shear zone forms a 100-200 m wide belt of relatively high penetrative strain, along the boundary between the Cameron River volcanic belt and the Sleepy Dragon Complex, and a family of discrete shear zones that decrease in number and density outwards on both sides of the belt over a total, but variable, width of 500 m. The zone of high strain tends to be wider in the volcanic belt than in the granitic terrane. At one locality 3 km north of Sleepy Dragon Lake, for example, moderately high strain prevails in the volcanics for about 120 m from the contact but, for only 10-20 m from the contact in the adjacent granitic gneisses, although discrete shear zones occur farther eastward in the gneiss. Foliation within the shear zone generally parallels the volcanic-granitic contact.

Although some description of the contact relationships have already been presented the main structural features are outlined here along with preliminary observations from continuing structural investigations along this contact (Kusky, 1986a, b, c, in press; R. Cullen, unpublished preliminary map, 1985).

Where the strain gradient is highest (increasing toward the volcanic-gneissic contact) volcanic rocks are so intensely deformed that primary features such as pillow lavas may be totally obliterated. Rocks on both sides of the contact have foliations parallel to the contact and steeply plunging stretching lineations. Steep lineations defined by elongation of pillow lavas, breccia fragments, and mineral lineations (hornblende or quartz-feldspar ribbons) are ubiquitous in the volcanic belt and generally most intensely developed in the high-strain zone.

Transposed layering, intensely folded quartz segregations, and vertically plunging sheath folds indicate very high strain in the gneisses (Kusky, 1986a, c.). Although the main foliation in the Sleepy Dragon boundary zone generally parallels the volcanic-granitoid contact, it cuts across older folded mylonitic foliations of the complexly deformed granitoid gneisses (Davidson, 1972; Kusky, 1986c).

Kinematic indicators in high strain zones in the gneisses (shear bands, rotated porphyroclasts, asymmetrical pressure shadows) as well as an overturned isoclinal fold in the pillow

lavas parallel to the volcanic-granite contact, all suggest that the Cameron River belt was thrust over the Sleepy Dragon Complex (Kusky, 1986c,d). Lenses, streaks and slivers of granitic gneiss that locally occur within highly strained amphibolite along the contact may have been tectonically incorporated into the high strain zone. This interpretation implies basement involvement during thrusting.

The volcanic belt has numerous lineaments that are sub-parallel to the trend of the belt or the eastern boundary but only the most prominent features are shown in Figures 2 and 3. Although some mark discrete shear zones (Fig. 8b) and others lithological contacts, or both, the significance of most has not been determined.

Discrete shear zones in the volcanic succession and within the Sleepy Dragon Complex, that parallel the volcanic-granitic contact, are inferred to have a common origin related to movement along the high-strain zone. If, as suspected, many lineaments within the belt are shear zones, it is possible that the volcanic belt is complicated by fault imbrications and some of the interfingering of the Webb Lake Andesite in the Cameron River Basalt could also be due to fault repetition, but this inference has not been documented. Kusky (1986a) has suggested that several and possibly tens of kilometres of cohesive, ductile movement are needed to explain the observed deformation in the Cameron River volcanics and the Sleepy Dragon Complex. No marker units, however, have been identified that could be used to document the amount of displacement that has taken place.

Amacher shear zone

This zone is a north-northeasterly trending, 1-4 km wide (Henderson, 1985) belt of mylonite bordering the eastern side of the Sleepy Dragon Complex in the Sunset Lake and Tumpline Lake subareas. Large amphibolite dykes follow the volcanic-granitic boundary west of Amacher Lake and near the north end of the belt. Granitic rocks, amphibolite dykes and volcanic rocks are all intensely deformed near this boundary, and high strain extends up to 1 km into the volcanic belt (See Sunset Lake Basalt). Commonly along the granitic-volcanic contact, lenses and discontinuous screens of amphibolite within the granitic mylonite produce a lensoid gneissic rock. Likewise, sheared amphibolite dykes along the volcanic boundary contain elongate bodies of granite.

Weathering along shear planes produces prominent lineaments in the granitic terrane parallel to the boundary of the volcanic belt that are clearly visible in air photographs. Almost all foliations in this mylonite zone dip steeply (75-85°). In contrast to the adjacent volcanic belt, mylonites in the granitoid shear zone have horizontal to very shallow plunging lineations and abundant evidence for dextral strike parallel movement expressed by c-s band structure (White, et al., 1980; Simpson and Schmid, 1983) and rotated megacrysts. Thus, structures in the Amacher shear zone are more consistent with dextral transcurrent motion than thrusting. The steep plunging stretching lineations, ubiquitous in the adjacent volcanic belt probably relate to a completely different movement picture.

Fault bounded lenses of Sleepy Dragon gneiss that locally cut obliquely across the volcanic belt northwest of Amacher Lake and are locally separated from each other by a dextral transcurrent fault, were emplaced later than the main deformation in the volcanic belt and may be related to movement in the Amacher shear zone.

Although the Amacher shear zone may be the site of relatively late transcurrent movement, it is still Archean as indicated by plutons of the Prosperous Granite which intrude the shear zone.

Payne shear zone

The Payne shear zone is a 1-2.5 km wide north west trending belt of mylonites, (Henderson, 1985, Map 1601A), exposed for a distance of 40 km across the northeastern corner of the area mapped. The zone occurs mainly in the Meander Lake Plutonic Suite but also cuts off and penetrates the volcanic belt west and south of Payne Lake. Deformation varies from mildly crushed zones in the plutons and extensive breccia zones containing quartz stockworks (Henderson, 1985) to dense ultramylonite. Where the shear zone overlaps the volcanic belt west and south of Payne Lake a mixed zone comprises 1) pods and screens of amphibolite within cataclastic and sheared granite, 2) highly strained amphibolite containing lensoid blocks and layers of granite to produce a mafic-felsic gneiss, and 3) granitic dykes that have intruded parallel to foliation. Tightly folded foliations in mixed zones show transposed layering, intensely folded felsic layers, and axial planes that are roughly parallel to the trend of regional foliation. Foliations within the shear zone are almost always steeply dipping (80-85°) toward the southwest or vertical, and rarely have northeasterly dips.

South of Payne Lake a 200 m wide zone of mylonite, between the Payne Lake Formation and adjacent granitic terrane, is a thin layered (1mm - 15 cm) succession of dark and light grey bands. Dark bands, which vary from even planar structures with sharp contacts to thin tapering lenses, comprise a myriad of microscopic lentils separated by fine mortar of crushed felsite and oriented muscovite whereas the light layers are granoblastic felsite. The mylonite, which lacks relics of coarse grains may be sheared aplite, felsite dykes or volcanics of the Payne Lake Formation.

South of Payne Lake the shear zone penetrates the volcanic belt for 1.5-2 km where the deformation is more heterogeneous, and localized discrete shear zones characterized by tight intrafolial folds commonly contain tortuously contorted lenses and veins of quartz.

The Payne shear zone may be younger than the shear belts that bound the Sleepy Dragon Complex. It cuts the post-volcanic, and in some places post fold deformation, Meander Lake Plutonic Suite, but is cut by the early Proterozoic faults in the area (Henderson, 1985).

Stratiform shear zones

Shear zones are common along contacts between contrasting lithologies within the supracrustal succession. Boundaries between rhyolite domes and mafic volcanics in the Cameron River and Sunset Lake subareas are invariably sites of brecciation and ductile deformation.

Contacts between Burwash sediments and volcanics of the Beaulieu Group vary from highly schistose, commonly marked by gossan zones, where the sediments contact massive or pillowed lavas, to conformable boundaries that do not show anomalous strain where the volcanic belts are layered clastic rocks.

The shear zones presumably result from strain gradients that develop where two bodies of contrasting rheologies deform together. This concept is most dramatically demonstrated in the carbonate unit that occurs between rhyolite and Burwash sediments near Lake 1052 and at the north end of Turnback Lake. Carbonate units are essentially deformed carbonate, impregnated rhyolite breccias and grits. Foliation, defined by strong preferred orientation of blocks, lenses and slivers of rhyolite, dips steeply parallel to boundaries of the unit and to foliation in adjacent volcanic rocks. High strain is indicated by boudins of rhyolite and pegmatite, highly convoluted layers and veins, extensive flattening of clasts to thin discs, fine comminution of rhyolitic material to sand size, and orientation of clasts of all sizes to form the prominent foliation.

Proterozoic faults

Faults and fractures of probable Proterozoic age include northerly and northeasterly trending major transcurrent faults, northwesterly trending fractures occupied by mafic and felsic dykes, and minor faults that displace late dykes.

Regionally extensive northerly trending faults (Fig. 51), in the Sunset Lake and Tumpline Lake subareas, show left-lateral movement: two faults have displacements of 1-4 km. Faults east and south of Sunset and Payne lakes have been traced for 80-100 km (Henderson, 1985). These faults displace volcanic belts and offset some granitic plutons. Southeast of Turnback Lake the east-southeasterly trending wings of the volcanic belt end abruptly against granitic rocks along the north-northeast-trending fault. South of Payne Lake, north- to northeasterly-trending faults cut obliquely across, and displace, the volcanic belt and the granitic terrane to the east. Some of these faults have a normal component that displaces the anticlinorial structure of the volcanic belt so as to effect an abrupt decrease in the width of the belt southwest of Payne Lake. Some northeasterly faults across the belt south of Payne Lake show dextral offsets. They may form a conjugate set with the northerly faults, but are less well developed.

Major movement on northerly trending faults took place after volcanism, some plutonism and deformation by folding. Some faults of this trend cut Dogrib dykes of Proterozoic age (Henderson, 1985).

North to northwesterly trending, fresh mafic dykes and felsic dykes define late sets of fractures that transect the Sleepy Dragon Complex, deformed volcanic belts and plutons. The fractures clearly postdate major regional structural patterns and therefore may be Proterozoic similar to the dykes that mark them.

Northeasterly trending faults that cause minor displacements of a northerly trending fresh diabase dyke, north of Sunset Lake, are the youngest faults known in the area.

SUMMARY AND EVOLUTION OF THE CAMERON-BEAULIEU VOLCANIC BELTS

This section summarizes conclusions and interpretations regarding the evolution, environments and style of volcanism represented in the Cameron-Beaulieu volcanic belts. In some respects the interpretations are idealized as a consequence of deciphering primary states through extreme deformation and metamorphism and numerous structural uncertainties.

Major gaps remain in structural, geochronological and isotopic data necessary to place constraints on tectonic processes. Accepting these limitations the physical volcanic scenario (illustrated by cartoons in Fig. 52, 53 and 54) is presented, followed by a chronicle of the main magmatic and structural events and a brief discussion of models that attempt to place the belts in a plausible tectonic framework.

Volcanism in the Cameron River subarea

The earliest volcanism in the Cameron River subarea was voluminous extrusion of basaltic lava as extensive subaqueous fissure eruptions (Cameron River Basalt). During each new eruption dykes intruded the preceding flows: some intruded laterally, as sills, before emerging as pillow lavas. Thus the basalt accumulated as a series of lava ridges and piles that built upwards and outwards, possibly to form an elongate shield or oceanic ridge. Within the volcanic pile lavas ponded against fault or constructional scarps which shed talus of pillow breccia and coarse basaltic volcanic clastic material. Submarine avalanches, possibly triggered by earthquakes, removed portions of the pile to create broad channels or valleys within the belt. Once formed, these depressions may have remained the locus of submarine erosion and deposition of volcanic debris presently represented by the thick lenses of volcanic clastic sediments within the Cameron River Basalt).

Mud and silt accumulated locally during pauses in lava effusions.

Rhyolitic magma erupted through the basaltic pile from at least two centres to form lava domes flanked by aprons of talus (Fig. 52b). During eruption and destruction of the domes, subaqueous debris and mass flows distributed detritus for 5-10 km from the domes into fault troughs and depressions in the mafic lavas and outward beyond the volcanic belt into the sedimentary basin.

Avalanching from fault scarps in the volcanic belt produced rubble of mixed rhyolite and basalt provenance.

Major eruptions of andesitic magma (the Webb Lake Andesite, Fig. 52c) followed rhyolitic volcanism. The change from basaltic to andesitic compositions was rapid, as evinced by the conformable relations with the preceding basalts and lack of intervening sediments, except for pillow breccias, which are common. The first andesitic effusions were pillowed and massive lavas that filled elongate valleys and fault troughs in the underlying basalts. These early lavas were fissure eruptions. Eventually, volcanism coalesced into central eruptions. Andesitic volcanism took place in the southern

part of the belt, probably while basaltic volcanism continued to the north.

The association of rhyolite immediately preceding andesitic volcanism suggests that following basaltic volcanism a magma source, possibly a vertically zoned chamber, was being tapped from which rhyolite differentiates were the first to erupt, followed by more mafic andesites, then finally, by a return to basalt (Dome Lake Basalt) at the end of the eruptive series. The rhyolite could represent purely the end product of fractional crystallization, or magma from a new source, such as melted sial, possibly present near the site where basaltic magma was introduced to the crust.

The general lack of granitic detritus or inclusions in the volcanic succession might suggest that the volcanic pile did not develop on the granitoid basement against which it presently makes contact. However, the small volume of granitic-volcanic conglomerate, quartz-feldspar arenites and quartzite of the Raquette Lake Formation, that interfingers with the volcanics near the place where the volcanic belt tapers out (Henderson, 1985), indicates proximity of a granitic source terrane at the time of volcanism.

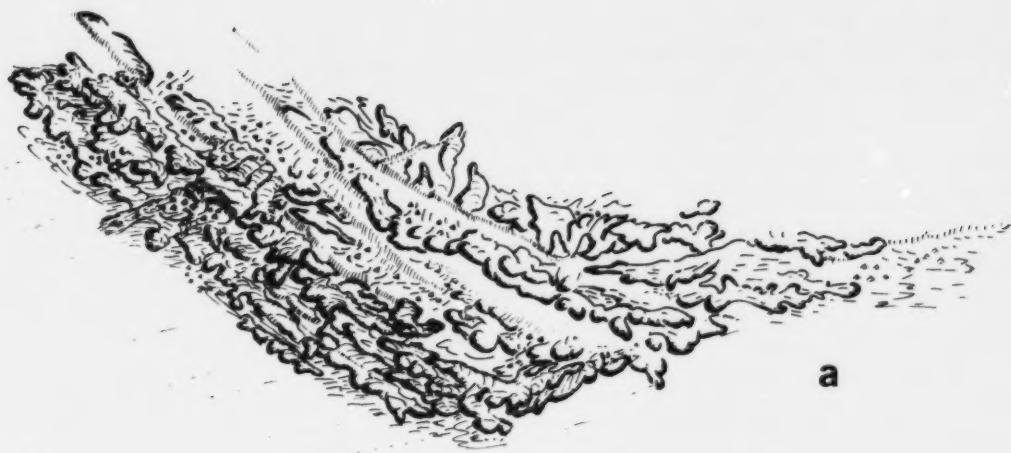
Speculation on eruptive centres, eruption type and paleomorphology

In the Cameron River subarea four features suggest proximity to eruptive centres: (1) dense swarms of mafic dykes that are contemporaneous with the lavas; (2) areas of local thickening within the Webb Lake Andesite; (3) formations of lava that occur as thick pods in only one locality (Dome Lake Basalt), and (4) rhyolite domes. Although most of the lava erupted from fissures, central eruptions dominated the last stages of volcanism.

Fissure eruptions

At least some of the dykes and sills filled fissures from which surface flows erupted. Evidence that the intrusions fed the

Figure 52. Volcanism in the Cameron River subarea. (a) Mafic magma first rose and effused from extensive linear fractures in submerged terrane of extended sialic crust. The first effusions probably formed elongate hills above their fissures. The basin subsided gradually as continuous voluminous effusion of mafic lavas built the volcanic pile upwards and outwards by the accumulation of a succession of overlapping pillowed flows (Cameron River Basalt). Syn-volcanic faults shed talus of pillow breccia within the pile and granitic-basaltic detritus at higher eastern levels of the basin margin. Local avalanches produced small scarps and possibly submarine channels in which coarse volcanic detritus accumulated. (b) Eruption of rhyolite domes and flows from centres marginal to, and on top of, the Cameron River Basalt. Debris flows from disintegrating domes distributed rhyolite and some mixed rhyolite and basalt detritus into valleys in the volcanic pile beneath, as well as into the adjacent sedimentary basin. (c) Eruption of andesitic magma (Webb Lake Andesite) produced pillowed flows in valleys and fault troughs in the Cameron River Basalt. Volcanism, at first from fissures, later coalesced into central eruptions which built a series of submarine mountains along the western edge of the volcanic belt.



flows include: (1) dykes locally pass into pillow lavas; (2) dykes and flows locally have similar distinctive petrographic features; (3) dykes and flows are similar chemically; (4) sills form an integral part of the volcanic stratigraphy; and (5) dykes profusely intrude the volcanic succession but not the younger Burwash sediments. These relations infer contemporaneity of the intrusions and extrusions. As would be expected, in the central part of the belt the concentration of dykes and sills tends to be highest in the stratigraphically older (eastern) parts of the belt. The dyke swarms suggest that lavas erupted from elongate vent systems over an area at least 10-20 km long and 3-5 km wide. This plumbing system, however, does not necessarily reflect the geometry of the magma chamber. The dykes may have been fed from one or more central conduits or reservoirs located at the roots of, or within, the deep fracture system rather than directly upward from long narrow zones in the mantle (as documented at Kilauea Volcano; Swanson et al., 1976; Duffield et al., 1982; and modelled by Fiske and Jackson, 1972). Similarly, in Iceland, although many fissure eruptions may have been fed vertically from subcrustal or mantle levels, the majority of fissure swarms near central volcanoes are fed laterally for distances up to 70 km from high-level magma reservoirs (Sigurdsson and Sparks, 1978; Einarsson and Brandsdottir, 1980; Sigurdsson, 1985). The parts of the belt where dykes are most abundant (i.e. near Webb Lake) probably were nearest the source chamber(s).

Central eruptions

The younger parts of the Webb Lake Andesite show features that suggest that it erupted from central vents rather than fissures. Locally, thick accumulations about 4 km long, that form major protuberances in the western side of the volcanic belt, thin abruptly to the north and south. In these areas the proportion of clastic rocks (pillow breccias, hyaloclastites, volcanic breccias) to pillow lavas generally increases upwards (westwards) in the stratigraphic section, and the formation shows more abrupt facies changes and much more varied stratigraphy than in other parts of the belt. Furthermore, the Webb Lake Andesite lacks the intrusive swarm that is characteristic of the Cameron River Basalt. Associated with the thickened parts of the Webb Lake Andesite are rhyolite domes and related breccias, and at least one local pile of basaltic effusives (the Dome Lake Basalt). These features are compatible with eruption from two or more central vents. The Dome Lake Basalt may represent an adventive eruption on the flank of a stratovolcano or it could be part of the Cameron River Basalt that was farther west than the andesites.

Thus, in the Cameron River belt the compositional change from basalt to andesite appears to have been accompanied by a major change in the style of volcanism, from extensive fissure eruptions to central vent eruptions. The eruptions may have coalesced along fracture zones into the few centres around which the last effusions accumulated.

Paleomorphology

In spite of the intense deformation and limitations of the two-dimensional view afforded by the present level of erosion, a reasonable assessment of the original morphology of the volcanoes can be made by taking into consideration the distribution, thickness and stratigraphy of formations in the belt in comparison with the features of recent volcanoes. During the early stages of volcanism, voluminous effusions from numerous fissures accumulated (Fig. 52a) to form an oblong pile of overlapping flows. This accumulation probably formed a series of low abyssal hills, pillow mounds or ridges, and tentacular pillowed flows.

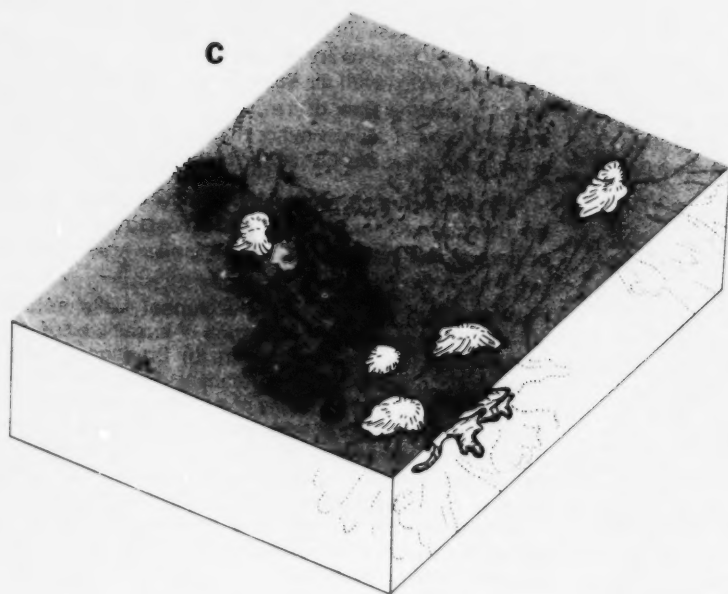
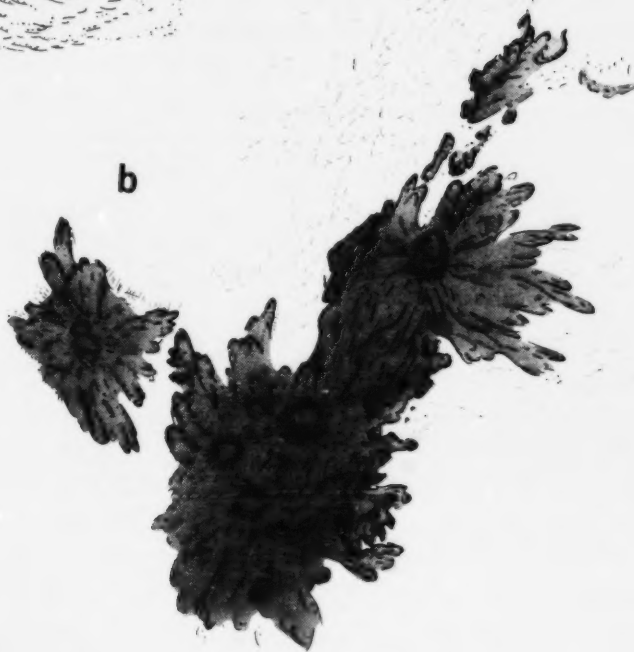
Synvolcanic faults produced numerous small scarps within the pile which were rapidly covered over by succeeding flows. Scarps and channels formed as a result of submarine landsliding.

Rhyolite domes that emerged mainly near the end of basaltic fissure eruptions probably were steep-sided mounds, in the order of 200-300 m high and 1000 m in diameter, that grew on top of, and at the margin of, the basaltic pile. They were flanked by lobes of breccia and distally thinning aprons of detritus. Tongues of rhyolitic debris accumulated in small grabens and topographic depressions between ridges in the mafic pillow lavas.

The inferred change in style of volcanism from fissure eruptions to central eruptions resulted in the growth of a complex submarine stratovolcano (the Webb Lake Andesite, and Dome Lake Basalt) that may have been 5 km across and had three, and possibly more, peaks 0.5-1 km in diameter. Coarse breccia in the Webb Lake Andesite represents subaqueous landslides on the slopes of submarine mountains.

The overall volcanic succession probably formed an elongate shield locally surmounted by submergent stratovolcanic cones. Although pillows in the belt generally face westward, the succession may never have been a vertically stacked pile of lavas. The present surface of erosion exposes a view, perpendicular to planes and axes of maximum strain, of a highly deformed agglomeration of originally overlapping lavas that may never have accumulated to any great thickness. Hence, the present belt reflects the width of a deformed and possibly tectonically imbricated and thickened lava succession rather than its original thickness.

Figure 53. Volcanism in the Tumpline Lake subarea. (a) Effusion of basaltic pillow lava (Tumpline Basalt) from major fracture systems. Debris flows carried detritus from the pile and synvolcanic scarps into the basin. (b) Eruption of rhyolite from several centres (Turnback Rhyolite and Sharrie Rhyolite). Explosive activity, either shallow submarine or subaerial, produced pyroclastic flows. Debris from these edifices dispersed widely, probably as sediment gravity flows, over the previously deposited basalts and into sedimentary basins beyond the volcanic pile where they formed the first volcanic deposits in that part of the basin. (c) Submarine effusion of basalt through part of the felsic pile deposited pillowed mounds and lenses on flanks and in depressions between felsic edifices. Emergence of felsic centres above sea level formed ephemeral islands. Sediments deposited in basins between islands.



Volcanism in the Tumpline Lake subarea

Volcanism in the Tumpline Lake subarea (Fig. 53) was strongly bimodal. In the northern areas (between Tumpline and Turnback lakes) volcanism began with the subaqueous effusions of basaltic pillow lavas. The basement against which the volcanics are now deformed may not have been at its present location at the time of volcanism. Submarine debris flows carried basaltic detritus from this major volcanic pile for 5-10 km, and possibly further, to the east and southeast. Most of this material was deposited before the succeeding rhyolite eruptions.

In the southern parts of the present subarea (south of Detour and Tumpline lakes) volcanism appears to have begun with voluminous eruption of rhyolite. Although it is possible that these rhyolites are underlain by basalts now concealed in the core of the large anticlinorium, and the mafic pile was considerably more extensive than the present exposure (as extrapolated in Fig. 53b), the present data suggest that rhyolite was the first volcanic product. This episode of rhyolite volcanism took place from two or three major centres stretching from the present Tumpline Lake to just south of Victory Lake. Eruption of the Sharrie Rhyolite began with explosive activity in a shallow marine environment. The explosions, possibly phreatomagmatic eruptions that broke the surface of the water, distributed pyroclastic material for several kilometres in the surrounding area, to be deposited as water-laid ash on sediments accumulating in the basin adjacent to the volcanic belt. Explosive volcanism was followed by quiet effusion of magma to form a complex of rhyolite flows and domes that possibly emerged as ephemeral islands.

In places south of Tumpline Lake, the simultaneous eruption of basalt and rhyolite from different vents may account for intertonguing of these rock types.

Basaltic magma erupted through the rhyolite complexes and effused as massive lavas on the flanks of subaerial edifices and rhyolite-block-bearing pillow lavas that partly filled the low-lying areas of submarine topography between or on the submerged flanks of rhyolitic edifices (Fig. 53c). Basaltic eruption in the southern parts of the area may have overlapped with the major early basaltic eruptions farther north. Amphibolite dykes, now folded within the early rhyolites, presumably mark the channelways through which the basaltic magma found its way to the surface.

Along the northern and eastern sides of this belt, voluminous rhyolitic volcanism, from at least three major centres postulated to be near Rex, Turnback and Tumpline lakes, succeeded the mafic effusions. The abundance of pyroclastic rocks indicates that volcanism was dominated by either shallow marine or subaerial explosive eruptions. The centre southwest of Tumpline Lake may have been a large, partially emergent edifice where subaerial pyroclastic flows were preserved. Presumably, widely dispersed ash, resulting from these explosive centres, settled at various times in the adjacent seas to become interbedded with mafic and felsic epiclastic debris that accumulated on the submerged flanks of the volcanoes. Deposition of subaqueous ash flows is likely, but such events have not been documented. Although some of the bedded rhyolites that prevail south of Tumpline Lake

may represent subaqueous pyroclastic flows, presumably much of the rhyolitic detritus was distributed into the basin by sediment gravity flows (Sigurdsson et al., 1980*; Schiener, 1974). Hence, part of the felsic pile probably extended beyond the limits of the underlying mafic pile to become the first volcanics to be deposited in that part of the basin.

Large amphibolite dykes within rhyolite northwest and southeast of Tumpline Lake suggest that late-stage basaltic volcanism may have been more common than indicated by the present stratigraphy. Near Consolation Lake, at the southern end of the subarea, a single amphibolite dyke has intruded Burwash sediments suggesting that mafic magmatism continued after the main volcanic formations were deposited.

Near the end of volcanism, the overall paleomorphology in this subarea probably consisted of a series of submarine ridges and seamounts, some of which emerged to form ephemeral islands.

The apparent high proportion of rhyolites to basalts in this subarea compared to that in the Cameron River belt may in part be a function of erosion. Perhaps the present surface exposes a much higher level within the volcanic pile, and thus preferential exposure of rhyolite that formed late in the volcanic history.

Eruptive centres

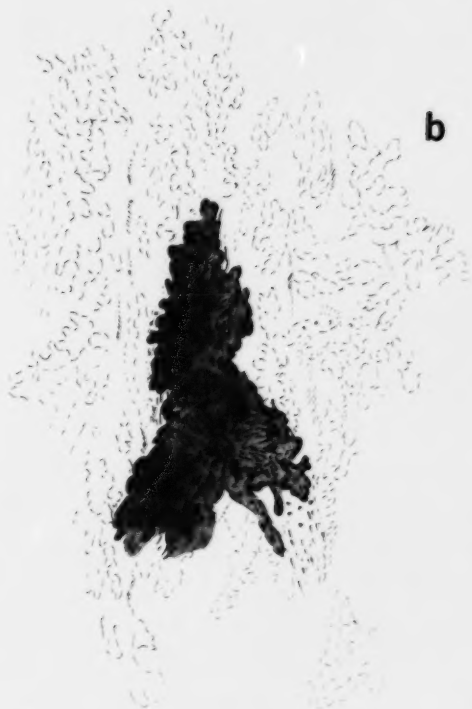
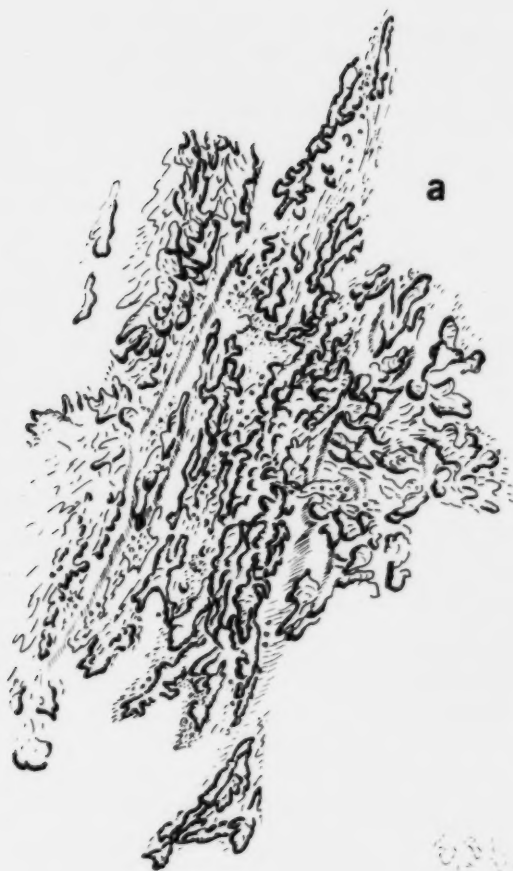
The striking bimodality of volcanic products in almost equal proportion that overlapped in time and space suggests that two distinct magma sources were being tapped contemporaneously, and the respective magmas followed the same structural paths to the surface.

The proportion of mafic magma erupted in the Tumpline Lake subarea is considerably less than in the Cameron River subarea. Accordingly, amphibolite intrusions, presumably related to the conduits of mafic eruptions, are relatively sparse. The intrusions are thick, large, sheet-like bodies that have been folded with the volcanic pile. Probably they represent horizontal flow of mafic magma derived from only one or two local centres rather than from extensive fissures as postulated for the Cameron River belt.

Proposed eruption centres or proximity to eruption centres for rhyolitic volcanics are areas that contain rhyolite flows or domes, and areas where the felsic piles are thickest and

Figure 54. Volcanism in the Sunset Lake subarea. (a) Voluminous submarine effusion of basaltic pillow lavas from fissures (Sunset Lake Basalt). (b) Eruption of andesite and dacitic magma (Alice Formation). Eruptions changed with time from quiet effusions of basaltic to andesitic lavas to explosive eruption of felsic andesite, dacite and rhyolite from central volcanoes. (c) Rise of rhyolite domes overlapped with previous intermediate to felsic eruptions. Some of these edifices and those of stage (b) emerged above sea level where welded tuffs were deposited. Erosion from rapidly built edifices produced extensive aprons of volcanic detritus (rhyolite rubble flanking domes and Payne Lake Formation in the southeastern part of the belt).

* Sigurdsson et al. stated that 70% of volcanogenic sediments in marine basins adjacent to the Lesser Antilles Arc were deposited in the form of sediment gravity flows; 40% of ash-fall is dispersed in the sediment; and only 20% of volcanic production remains on the volcanic islands.



contain abundant massive pyroclastic rocks. These areas are generally at, or near, the sites that are interpreted to have been subaerial by virtue of (1) presence of ash-flow tuffs that have characteristics of subaerial ash flows, (2) absence of intercalated sediments with volcanoclastic rocks, and (3) local occurrences of fluvial sediments (conglomerates). Subaqueous deposits flanking or distal to the eruptive centres commonly comprise interbedded rhyolite volcanoclastics with mafic pillow lavas or interbedded epiclastic deposits of rhyolite with mafic volcanic sands. Thin rhyolite units, up to 10 km from the proposed sources, were probably deposited as subaqueous mass flows issuing from the main piles, and in some places were water-laid crystal tuffs now interbedded with mafic volcanic sediments. Figures 53b and c show the proposed centres of rhyolitic volcanism, a proposed pattern of dispersal of pyroclastic and epiclastic debris from them, and the location of ephemeral islands inferred to have existed near the end of volcanism.

Volcanism in the Sunset Lake subarea

Volcanism began with voluminous subaqueous effusion of basalt pillow lavas. Figure 54 depicts the inferred paleomorphology and the sequence of events that developed this volcanic belt. The lavas probably formed an extensive elongate hummocky field of low hills and intervening valleys reflecting an underlying fracture-fissure system. Depressions in this topography accepted detritus derived from the accumulating pile. The southern edge of this field may have ended abruptly in a steep frontal slope, in some places, whereas in others it blended gradually with a tapering apron of volcanic debris sloughing from the volcanic pile.

Magma compositions became more felsic with time and eruptive products passed from basalt to andesite without any major break in the eruptive activity.

A change in the style and environment of eruption (Fig. 54b) accompanied this temporal change in magma composition. The mode of eruption changed from quiet effusion of basaltic and andesitic lavas to explosive eruption of felsic andesite, dacite and rhyolite. These volcanic products built submarine mountains, some of which emerged above sea level, where volcanism culminated with explosive subaerial eruptions. Rhyolite erupted during the final stages of volcanism from at least three main centres.

The waning stages of volcanism saw the rise of felsic domes on these edifices. At some centres effusions of intermediate to felsic compositions overlapped and interfingered. One dome complex (south of Sunset Lake) records a compositional transition from andesite through dacite to rhyolite. North of Sunset Lake, rhyolite volcanism began with early explosive activity followed by at least two, and probably more, episodes of dome growth. Near the south end of the lake, part of the felsic pile emerged above sea level where explosive eruptions deposited welded ash-flow tuffs.

Erosion degraded the stratovolcanoes of this composite volcanic pile and produced submarine fans that formed an extensive wedge of volcanic detritus south of the present Payne Lake.

Deposition of the Burwash sediments mainly followed volcanism in all three subareas. No stratigraphic or structural unconformities between the Burwash Formation and the Beaulieu Group have been recognized to indicate a gap in time between volcanism and sedimentation. The sediments were derived from exposed parts of the immediate volcanic pile as well as from a granitic terrane. According to Henderson (1985) the abundance of felsic grains in the greywackes suggests that these exposed parts included major felsic volcanic edifices possibly representing volumes as great as the volcanics now preserved at the present basin margins. That some sedimentation preceded volcanism, however, is suggested by iron-formation, metapelites and granitic cobble-bearing conglomerate at, or near, the base of the volcanic succession along the boundary zone with the Sleepy Dragon Complex. Furthermore, the turbidites directly contacting the Sleepy Dragon Complex all along its southwestern side, locally lie between the volcanic succession and the granitoid terrane and may underlie part of the volcanic belt south of Detour Lake. No major facies of coarse clastics and fan-glomerates typical of continental rifts are present. Although the sparse conglomeratic and quartz-rich units may have been deposited in debris or sediment gravity flows from distal sources, the conglomerates bearing both volcanic and granitic clasts may have been derived from source terranes, possibly fault scarps, where both rock types were exposed within the volcanic fields.

Chronicle of structural and magmatic events

The sequence of structural and magmatic events in the southwestern part of the Slave Province, where the Cameron River and Beaulieu River volcanic belts developed, involved: (1) extension of submerged sialic crust, emplacement of mafic dykes, volcanism, and diachronous sedimentation; (2) deformation of supracrustals involving horizontal shortening; (3) metamorphism that peaked during major fold deformation; (4) emplacement of granitic plutons and uplift of granitic basement into deforming supracrustals; (5) mild regional shortening and folding; (6) emplacement of Proterozoic dykes; and (7) wrench faulting that postdated all the previous volcanic and structural events. Figure 55 shows the general timing of the Archean events.

Profuse mafic dyke swarms indicate that extension took place both in the Sleepy Dragon Complex and beneath the developing volcanic belts. Although it has not been established unequivocally that the volcanic belts developed on, or against, the present Sleepy Dragon basement, or whether the two terranes have been tectonically juxtaposed, mafic magmatism did take place through sialic crust, now represented by the Sleepy Dragon Complex, which probably was basement to some volcanism even if it was not that of the presently adjacent volcanic belts. The amphibolite dykes within the gneissic terrane could well be coeval with those in the volcanic pile. Amphibolite dyke swarms within the volcanic belts are contemporaneous with volcanism and pre-date sedimentation which deposited most of the Burwash Formation.

Thus the preferred interpretation is that volcanoes erupted through submerged sialic crust and possibly also on adjacent

oceanic seafloor in one or more major areas now represented by the Cameron River and Beaulieu River belts and their extension to the Beniah Lake belt to the north. Sedimentation was diachronous, spanning and continuing long after volcanism.

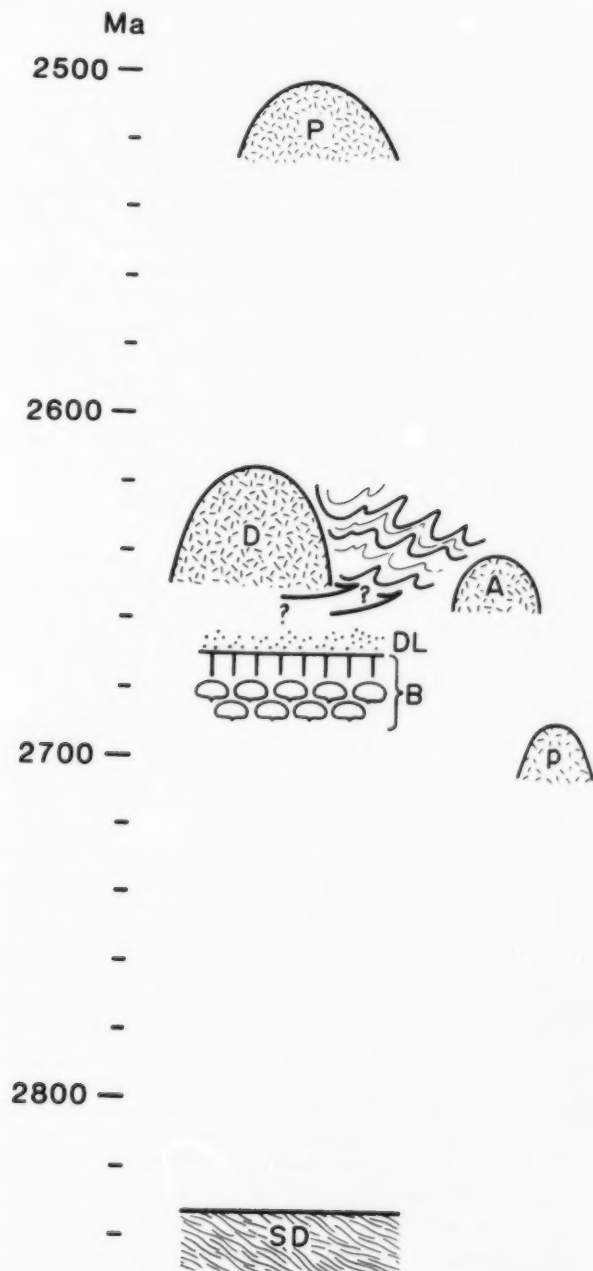


Figure 55. Generalized chronology of events near the Cameron and Beaulieu volcanic belts (see Geochronology section for source of radiometric dates). SD = Sleepy Dragon Complex, D = Defeat Plutonic Suite, P = Prosperous Granite, A = Amacher Granite, p = pluton in Sleepy Dragon Complex, B = Beaulieu Group, DL = Duncan Lake Group.

Deformation

Horizontal shortening throughout the volcano-sedimentary basin produced large-scale steep isoclinal folds and intense strain along boundaries between volcanics and basement. In the Tumpline Lake subarea, high strain and polyphase deformation produced complex fold interference patterns in the volcanic belt (Fig. 56). Large-scale isoclinal folds and anticlinoria moulded around the Sleepy Dragon Complex. In the Sunset Lake subarea the volcanic belt deformed into a broad anticlinorium and synclinorium. The Amacher Granite, which intruded the pile either before or during deformation, may have controlled the position of the anticlinorium. This 2644 ± 14 Ma (Zircon age; J.B. Henderson and O. van Breemen, pers. comm., 1986), premetamorphic pluton constrains the maximum age of deformation. In the Tumpline Lake and Sunset Lake subareas some deformation of volcanics may be related to diapiric emplacement of plutons either before or during deformation.

The apparent structural and stratigraphic simplicity of the Cameron River belt is deceiving. It is proposed that the belt deformed against the Sleepy Dragon Complex and that high strain produced pseudoconformable relations within the volcanic belt and between the eastern side of the volcanic belt and the granitic basement. Thus the present volcanic assemblage represents a highly deformed belt that may be tectonically thinned or possibly imbricated and thickened.

The three- and possibly four-phases of folding and extreme diversity of deformation patterns displayed in the Cameron-Beaulieu volcanic belts require variable stress fields and their response to variable boundary conditions imposed by the basement and interaction with pre- to syndeformational granitic plutons. That the main deformation is Archean as indicated by ca. 2520 Ma plutons (Stockwell, 1962; Green et al., 1968; Green and Baadsgaard, 1971) of the unmetamor-

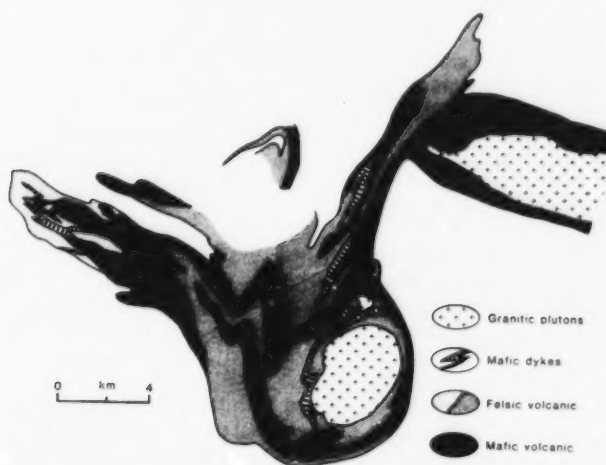


Figure 56. Polydeformed volcanic belt in the southern part of the Tumpline Lake subarea. The interpretation projects units through lakes and the Prosperous Granite, and shows the Mafic units as a highly strained essentially single unit that interdigitates with rhyolite.

phosed Prosperous Lake granite that cut discordantly across the southwestern end of the Beaulieu River belt, the Sleepy Dragon Complex and the Burwash sediments. High temperature-low pressure metamorphism accompanied plutonism and the main phase of deformation (Henderson, 1985; Fyson, 1980).

Late folds that show a consistent pattern in trends of axial plane foliations, regardless of their position with respect to basement, plutons and previous fold structures, record a strain of regional nature that apparently affected supracrustal rocks across the southern part of the Slave Province. This strain may have been synchronous with post-Archean wrench faulting.

The Sleepy Dragon Complex is bounded by shear zones and zones of high strain both in the granitic gneisses and the volcanic belts. Near vertical stretching lineations and kinematic indicators along boundary zones in both the Cameron River and Sunset Lake subareas, as well as fault-bound slabs of basement incorporated into the Sunset Lake Basalt, are compatible with thrusting of the volcanic belts over the Sleepy Dragon Complex (Kusky, 1986a,b; Lambert and van Staal, 1987). In the Cameron River subarea, however, the irregular angular boundary of the Sleepy Dragon Complex does not have the appearance of a simple thrust plane. Faults in the basement that displace the contact do not seem to penetrate the volcanic belt for any appreciable distance and are not reflected in the western boundary of the volcanic belt. This irregularity could be a relic of an original faulted topography against which the volcanic belt was deformed. Unresolved questions that are critical to resolving this problem are:

1. What is the precise relationship of the shear zones to the basement irregularity?
2. Does a basal décollement transect or wrap around the angular boundary?
3. At what time in the structural evolution did the irregularity form?

The thrust scenario would infer that movement zones had shallow to moderate attitudes which subsequently must have been steepened. It has not been determined if the sense of movement is consistent all around the Sleepy Dragon Complex with thrusting, and if so what mechanism then steepened the lineations to near vertical all around the basement terrane.

A further complication is that granitic gneisses in the Amacher shear zone have horizontal lineations and movement indicators suggesting dextral transcurrent movement along part of the Sleepy Dragon boundary.

The structural history in the boundary zone is obviously complex and cannot be resolved with the present data. As a first approximation it is proposed that movement and deformation in the granite-greenstone boundary zones could be related to a combination of effects such as: (1) a strain gradient expected where two bodies of contrasting rheology are deformed together; (2) thrusting of the volcanic belt over the basement; (3) uplift of the Sleepy Dragon Complex; and (4) transcurrent movement along some boundary zones. An original irregularity along the western side of the Sleepy

Dragon Complex might be preserved if the main décollement during thrusting was not at the exact contact and that movement took place along a family of imbricate shear zones within both the volcanic and granitic terranes.

Plutonism

Emplacement of plutons probably spanned a time from the late stages of volcanism to post-deformation by folding. Those that intruded the volcanic belts have characteristics of bodies emplaced at both intermediate and high levels in the Earth's crust. The north-northeasterly elongation and distribution of many plutons in the Sunset Lake and Tumpline Lake subareas suggest emplacement along structural lines related in some way to the boundary of the Sleepy Dragon Complex.

Some of the earliest plutons may have arched or domed parts of the volcanic pile during late stages of volcanism and possibly contributed locally to emergence of volcanic piles above sea level. Plutons of the Defeat granodiorite in the Tumpline Lake and Sunset Lake subareas, have semi-concordant boundaries with sediments and the volcanics, and may predate the main folding (in part) but continued to rise diapirically during deformation. As deformation continued, the supracrustals became progressively more tightly moulded around them. The elongate Amacher Granite in the Sunset Lake subarea appears to have controlled the position of an anticlinorium in the volcanic belt. Late post-deformational plutons that cut sharply across all major fold structures (Prosperous Granite) and possibly parts of the Defeat and Meander Lake suites represent major plutonic episodes that occurred some 40+ Ma later than formation of the greenstone belt.

The highly deformed sediments and volcanics do not show a consistent regional pattern. The diverse patterns must reflect diverse stress fields and boundary conditions. Diapiric emplacement of granitic plutons (Drury, 1977) may have played a role in producing the divergent strain patterns in the volcanic belts. Although no conclusive studies on diapirism in the Slave Province have been made, many features of plutons and surrounding sediments are consistent with this mode of emplacement. Maps by Henderson (1976; 1985) and McGlynn (1977) indicate that at the present level of erosion plutons occupy about 25 % by area of the Yellowknife basin, and near the Cameron-Beaulieu volcanic belts, pre- and syn- deformational plutons occupy major proportions of the adjacent sedimentary basin as well as the basement complexes (Fig. 38). Structural patterns in the supracrustals clearly relate to those plutons as well as to irregular basement blocks, such as the Sleepy Dragon Complex, against which the volcanics deformed. Possibly, rising granitic diapirs not only domed and steepened attitudes in sediments and volcanics, but also inflated basement blocks and contributed to their uprise. Considering the theoretical plausibility of diapiric emplacement of granitic plutons (Marsh, 1982, 1984), its implied importance in producing deformation in some volcanic arcs (Tobish et al., 1986), and the clear relationship of plutons to structural patterns in the southern Slave Province, it is unlikely that plutons, except a few, were merely passively emplaced. It is proposed that during episodes of voluminous granitic magmatism, diapirically rising plutons

played an important role in modifying the deformational pattern externally imposed on the Yellowknife supracrustal basin by tectonic events. Hence, major folding in the basin progressed over a period of time that spanned numerous plutonic and concomitant structural events in various parts of the basin.

Tectonic models

The first models for the evolution of the Yellowknife Supergroup (McGlynn and Henderson, 1970; Lambert, 1977; Henderson, 1981, 1985) proposed that regional extension in a ca. 3 billion year old granitic basement, produced horsts and grabens in which Archean supracrustal rocks accumulated over a period of 10 to 15 million years, about 2670 million years ago. Marine inundation followed initial stages of intracratonic rifting. Volcanoes erupted along the faulted basin margins and subsequently the margins shed granitic and volcanic debris into the evolving basin. Rifting never extended beyond the graben stage as the basins were presumed to be floored by granitic basement and there is no evidence that the pre-Yellowknife crust was ever completely rifted such that oceanic crust formed within the basins. The Burwash turbidites are seen as submarine fans that extended into the basin from the margins.

The model provides a logical framework of crustal fracturing followed by volcanism, subsidence and erosion to produce the sedimentary basins, and it accounts for the ubiquitous occurrence of volcanics at the angular boundary between sediments and basement complexes. Criticisms of this model are that it does not appear to account for the general lack of terrestrially derived coarse elastic successions characteristic of rift sequences (Ziegler, 1982; Burke et al., 1985), the lack of expected facies changes in the sediments as they approach the granitic basement, the absence of alkaline magmatism, nor does it provide a dynamic framework to account for the polyphase deformation and horizontal shortening ubiquitous in the supracrustal rocks. Although the model requires widespread granitic basement across the Slave Province, basement to the supracrustals has been documented only at a few widely-spaced localities (Baragar and McGlynn, 1976).

A modified version of the rift model may apply to part of the Slave Province, but it is too simple a scenario for the province as a whole. Consequently these earlier ideas have been followed by models suggesting that the greenstone belts of the Slave Province formed in back-arc basins (Kusky, 1986a 1987; Helmstaedt and Padgham, 1986), or represent parts of accretionary complexes overlain by trench turbidites (Hoffman, 1986, Kusky 1986a; Fyson and Helmstaedt, 1987).

Tarney et al. (1976) first proposed that marginal or back-arc basins provide an actualistic counterpart to Archean greenstone belts using the analogy of the Sarmiento complex of southern Chile. Similarly, the supracrustal succession of the Yellowknife basin has been compared to this complex (Helmstaedt and Padgham, 1986) and to the Miocene Fossa Magna of Japan (Folinsbee et al., 1968). The attraction of this model is that it is compatible with the geochemistry of the volcanic rocks (i.e. tholeiitic basalts followed by calc-

alkaline andesites) and accounts for the generation of large volumes of post-volcanic, calc-alkaline plutons.

In the Cameron-Beaulieu volcanic belts, however, the ophiolitic part of this analogy is not adequately represented. A major part of the typical ophiolite sequence (Constantinou, 1980), the sheeted dykes, and layered gabbroic plutons are missing, and ultramafic rocks, although present, form sparse fault-bound slivers within boundary shear zones and could be dykes. Although there are dense mafic dyke swarms, none form a continuous layer of 100% dykes as in true sheeted dyke complexes (Baragar et al., 1987; Gass, 1980) and the densest swarms intrude granitic gneisses. Dyke swarms as dense as those in the Cameron-Beaulieu belts prevail in high standing volcanic edifices such as the Koolau complex in Hawaii (Walker, 1986) and are probably a common feature at critical levels of pressure and stress within major volcanic edifices or areas of voluminous basaltic eruption.

The most comprehensive tectonic model yet proposed for the origin of granite-greenstone terranes is that of crustal accretion which visualizes greenstone belts as remnants of fore-arc accretionary complexes where the volcanic and sub-volcanic rocks are allochthonous island arcs, aseismic ridges, submarine plateaus and sea mounts. The juxtaposed island arcs and bathymetric highs were delaminated from subducting oceanic lithosphere and overlain by allochthonous pelagic and deep sea fan conglomerates and trench turbidites (Hoffman, 1986, Kusky 1986a) in a prograding arc-trench system. Granitoids coeval with calc-alkaline volcanism are interpreted as the accreted roots of island arcs. Older basement may represent accreted microcontinents possibly including rift basins related to their fragmentation or active continental margins rifted during back-arc spreading. The model accommodates a wide variety of initial environments, and acknowledges that some greenstone belts were formed on older continental crust but have subsequently been detached.

Undoubtedly these models will be tested and modified as more structural, isotopic, and geochronological data and paleoenvironmental interpretations are obtained.

Models for the evolution of the Cameron and Beaulieu volcanic belts are constrained by the following conclusions from this study.

1. Granitic crust was present at the time of volcanism. This is documented by foliated granitic boulders in conglomerates within basal parts of the volcanic succession; dense mafic dyke swarms in basement gneisses which are about 120 Ma older than the volcanic rocks; and the high proportion of rhyolite in the Tumpline Lake subarea that would seem to require melting of sialic crust for its derivation.
2. The volcanoes erupted mainly in a subaqueous environment.
3. Boundaries between the Sleepy Dragon Complex and the volcanic belts are shear zones or complex zones of high strain.
4. The sedimentary basin is not presently floored by a mafic layer (McGrath et al., 1983). It could be floored by granitic crust, of unknown age relative to the volcanics.

It would appear that the Cameron and Beaulieu volcanic belts developed in an environment of submerged, extended, and possibly block faulted, continental crust. The rifting model should not be ruled out entirely as a possibility during the early stages of development of these belts. The general lack of coarse clastic sediments is a feature of volcanic dominated rift basins where voluminous early volcanism infills the developing depressions nearly as fast as they subside (e.g. Kenya Rift: Cohen et al., 1986; Baker, 1986), that is where volcanic extrusion rates are high relative to the rate of rift extension. The Sleepy Dragon Complex may be a remnant of a passive margin that was once part of an intracontinental rift, but was subsequently subjected to horizontal tectonic regimes whereby the supracrustal rocks were severely deformed and transported onto the cratonic basement.

The volcanic belts and the Sleepy Dragon terrane are but a fragment of a large and complex tectonic picture which could be accommodated in either the back-arc basin or accretionary models. Assessment of the local tectonic setting, however, depends on critical observations of the granite-greenstone boundary relationships which have not been adequately determined in this study.

Magma origin and emplacement

There does not seem to be a direct analogue of the totally subalkaline volcanic assemblage of these greenstone belts (tholeiitic lavas, calc-alkaline andesite, dacite, rhyolite) to assemblages in Phanerozoic or modern cratonic rifts (characterized by alkaline magmatism). Volcanic assemblages in the Slave Province as a whole bear a similarity in part to those at convergent plate margins in the Phanerozoic (Condie, 1982; Condie and Hunter, 1976) and to oceanic ridge tholeiites. In the Cameron River belt, where the ratio of tholeiitic to calc-alkaline compositions is high and sialic end members are minimal, magmas are presumed to be derived from the mantle.

Mantle derivation is more difficult to justify, however, in the Tumpline Lake and Sunset Lake subareas where the proportion of rocks having calc-alkaline composition is much higher, and the calc-alkaline rocks are dominantly dacites and rhyolites. In these belts, melting of the sialic crust is a more attractive model. Large amounts of mafic magma invading sialic crust could induce sufficient heating to cause melting and produce significant amounts of anatectic melts (Burnham, 1979; Marsh, 1984).

At the present level of erosion, mafic magma moved in extensive fissure systems (individual intrusions up to 6 km long), whereas felsic magma seems to have erupted from central sources. Not all dykes necessarily reached the surface, nor did they necessarily extend vertically downwards to their source. At some level in the crust the pressure in chambers of magma accumulation may be relieved by longitudinal flow along fractures. In Hawaii, Vogt (1976) and Lonsdale and Spiess (1979) described a model whereby a small number of vertical pipes feed magma horizontally to form elongate magma chambers. Such a mechanism is visualized for the mafic volcanism in the Tumpline Lake subarea and probably was prominent in the Cameron River subarea.

It is reasonable to assume that the intensity of dyking increases downwards so that a few dykes near the surface may gradually become a wider, more profuse swarm at depth. This geometry fits the thermal requirements and the tapered dyke complex modelled by Rudman and Epp (1983) in Hawaii. If the volcanic belts are allochthonous as implied in the structural interpretation, the dyke swarms within the belts are probably not rooted beneath their present locations and thus removed from their parent magma chambers.

However contrived, accumulation of mafic magma into chambers at the base of, or within the sialic crust, may have enough heat to cause melting and generate the calc-alkaline and felsic magma which differentiated and erupted mainly after a profuse effusion of basaltic lavas. Felsic magma chambers presumably represent higher level accumulations of crustal melt. Abrupt changes from basalts to rhyolites and from tholeiitic to calc-alkaline suites without interruption in volcanism suggest tapping of different contemporaneous magma sources.

Timing of volcanism

The Cameron River and Beaulieu River volcanic belts comprise numerous deformed volcanoes that erupted through, or along, the margins of slightly older granitoid terrane for a continuous distance of 160 km in the present map area, but extend farther to the north and east to make a total distance of 430 km. The high level of volcanic activity presumably relates to a high rate of crustal extension in these areas. The estimated original volume of the volcanic belts in the present map area is between 24 000 and 45 000 km³. This estimate is necessarily crude because of the numerous unknowns such as the original depth, extent and shape of the volcanic fields. Reasonable estimates for these unknowns fall well within the model determined by McGrath et al. (1983) from the Yellowknife volcanic belt based on interpretation of a gravity profile. They suggested that the volcanic belt at Yellowknife is between 3 and 5 km thick, less than 15 km wide, and roughly wedge-shaped in profile, tapering towards a sedimentary basin.

The 24 000 km³ value was determined by extrapolating units and form of the volcanic belts based on the present stratigraphic record. Estimates attempted to take into account the effects of polyphase deformation, shortening of the stratigraphy, and the amount of the volcanic succession removed by erosion. The 45 000 km³ estimate uses proportions similar to the model of McGrath et al. (1983).

If the rate of magma supply to the volcanic belts was within the ranges calculated for Hawaii (Swanson, 1972, calculated the current rate at Mauna Loa and Kilauea, considered a maximum, of 10⁻¹ km³.a⁻¹; Shaw, 1973, determined a rate averaged over the past 6 Ma of 3.3 x 10⁻² km³.a⁻¹) or Iceland (rate over the last 10 000 years is 4-4.8 x 10⁻² km³.a⁻¹; Thorarinnsson, 1967; Jakobsson, 1972), the volume estimated in the Cameron River and Beaulieu River volcanic belts could have accumulated within 50-100 000 a, assuming a constant supply rate. Lack of any recognizable time breaks in stratigraphy suggests that volcanism was more or less continuous for a time, then ceased. Even if the volume estimates are several times too low, the volcanic belts could have developed in less than 1 Ma.

REFERENCES

- Anderson, J.E.T.**
1969: Development of snowflake texture in welded tuffs, Davis Mountains, Texas; *Geological Society of America, Bulletin*, v. 80, p. 2075-2086.
- Autran, A., Fontelles, M., and Guitard, G.**
1970: Relations entre les intrusions de granitoides, l'anatexis et le métamorphisme régional considérées principalement du point de vue du rôle de l'eau: cas de la chaîne hercynienne des Pyrénées orientales; *Société géologique de France, Bulletin*, v. 7, XII, no. 4, p. 673-731.
- Baker, B.H.**
1986: Tectonics and volcanism of the southern Kenya rift valley and its influence on rift sedimentation; in *Sedimentation in the African Rifts*, ed. L.E. Fostick, R.W. Renault, I. Ried, and J.J. Tiercelin; Geological Society of London, Special Publication No. 25, p.45-57.
- Baragar, W.R.A.**
1966: Geochemistry of the Yellowknife volcanic rocks; *Canadian Journal of Earth Sciences*, v. 3, p. 9-30.
- Baragar, W.R.A. and McGlynn, J.C.**
1976: Early Archean basement in the Canadian Shield: a review of the evidence; *Geological Survey of Canada, Paper* 76-14.
- Baragar, W.R.A., Lambert, M.B., Baglow, N., and Gibson, I.L.**
1987: Sheeted dykes of the Troodos ophiolite, Cyprus; in *Mafic Dyke Swarms*, ed. H.C. Halls and W.F. Fahrig; Geological Association of Canada, Special Paper 34, p. 257-272.
- Bibikova, E., Baadsgaard, H., and Folinsbee, R.E.**
— U-Pb and Sm-Nd dating of gneissic tonalite boulders from a diatreme in the Con Mine, Yellowknife, NWT, Canada, with South African comparisons; *Terra Cognita*. (in press)
- Bowering, S.A. and van Schmus, W.R.**
1984: U-Pb zircon constraints on evolution of Wopmay Orogen, N.W.T.; in *Geological Association of Canada, Mineralogical Association of Canada, Canadian Geophysical Union, Joint Annual Meeting, London 1984, Program with Abstracts* v. 9, p. 47.
- Buddington, A.F.**
1959: Granite emplacement with special reference to North America; *Geological Society of America, Bulletin*, v. 70, p. 671-747.
- Burke, K., Kidd, W.S.F., and Kusky, T.**
1985: Is the Ventersdorp Rift System of Southern Africa related to a continental collision between the Kaapvaal and Zimbabwe cratons at 2.64 Ga ago?; *Tectonophysics*, v. 115, p. 1-24.
- Burnham, C.W.**
1979: Magmas and hydrothermal fluids; in *Geochemistry of Hydrothermal Ore Deposits*, second edition, ed. H.L. Barnes; John Wiley and Sons, p. 71-136.
- Burwash, R.A. and Baadsgaard, H.**
1962: Yellowknife-Nonacho age and structural relations; in *The Tectonics of the Canadian Shield*, ed. J.S. Stevenson; Royal Society of Canada, Special Publication No. 4, p. 22-29.
- Burwash, R.A., Baadsgaard, H., Campbell, F.A., Cumming, G.L., and Frolinsbee, R.E.**
1963: Potassium-argon dates of diabase dyke systems, District of Mackenzie, N.W.T.; *Canadian Institute of Mining and Metallurgy, Transactions*, v. 66, p. 303-307.
- Cann, J.R.**
1970: Rb, Sr, Y, Zr, Nb in some ocean-floor basaltic rocks; *Earth and Planetary Science Letters*, v. 10, p. 7-11.
- Carlisle, D.**
1963: Pillow breccias and their aquagene tuffs, Quadra Island, British Columbia; *Journal of Geology*, v. 71, p. 48-71.
- Cohen, A.S., Ferguson, D.S., Gram, P.M., Hubler, S.L., and Sims, K.W.**
1986: The distribution of coarse-grained sediments in modern Lake Turkana, Kenya: implications for clastic sedimentation models of rift lakes; in *Sedimentation in the African Rifts*, ed. L.E. Fostick, R.W. Renault, I. Ried, and J.J. Tiercelin; Geological Society of London, Special Publication No. 25, p. 127-139.
- Condie, K.C.**
1982: Early and middle Proterozoic supracrustal successions and their tectonic settings; *American Journal of Science*, v. 282, p. 341-357.
- Condie, K.C. and Hunter, D.R.**
1976: Trace element geochemistry of Archean granitic rocks from the Barberton region, South Africa; *Earth and Planetary Science Letters*, v. 29, p. 389-400.
- Condie, K.C., Viljoen, M.J., and Kable, E.J.D.**
1977: Effects of alteration on element distributions in Archean tholeiites from the Barberton greenstone belt, South Africa; *Contributions to Mineralogy and Petrology*, v. 64, p. 75-89.
- Constantinou, G.**
1980: Metallogenesis associated with the Troodos ophiolite; in *Ophiolites: Proceedings of the International Ophiolite Symposium, 1979*, Geological Survey Department, Ministry of Agriculture and Natural Resources, Cyprus, p. 663-674.
- Culshaw, N.G.**
1986: Proterozoic basement tectonics along the western edge of the Theon Tectonic Zone (N.W.T.) - compressional and transcurrent structures - a result of oblique convergence?; in *Geological Association of Canada, Mineralogical Association of Canada, Canadian Geophysical Union, Joint Annual Meeting, Ottawa 1986, Program with Abstracts*, v. 11, p. 60.
- Cumming, G.L. and Tsong, F.**
1975: Variations in the isotopic composition of volatilized lead and the age of the Western Granodiorite, Yellowknife, Northwest Territories; *Canadian Journal of Earth Sciences*, v. 12, p. 558-573.
- Davidson, A.**
1972: Granite studies in the Slave Province; in *Report of Activities, Part A*, Geological Survey of Canada, Paper 72-1A, p. 109-115.
- Dekker, A.G.D.**
1978: Amphiboles and their host rocks in the high grade metamorphic Precambrian of Rogaland/Vest-Agder, Southwest Norway; *Geologica Ultraiectina*, no. 17, 277 p.
- Dixon, J.M.**
1975: Finite strain and progressive deformation of models of diapiric structures; *Tectonophysics*, v. 28, p. 89-124.
- Drury, S.A.**
1977: Structures induced by granitic diapirs in the Archean greenstone belt at Yellowknife, Canada: implications for Archean geotectonics; *Journal of Geology*, v. 85, p. 345-358.
- Duffield, W.A., Christiansen, R.L., Koyanagi, R.Y., and Peterson, D.W.**
1982: Storage, migration, and eruption of magma at Kilauea Volcano, Hawaii, 1971-1972; *Journal of Volcanology and Geothermal Research*, v. 13, p. 273-307.
- Einarsson, P. and Brandsdottir**
1980: Seismological evidence for lateral magma intrusion during the July 1978 deflation of the Krafla volcano in NE Iceland; *Journal of Geophysics*, v. 47, p. 160-165.
- Fiske, R.S.**
1963: Subaqueous pyroclastic flows in the Ohanapecosh formation, Washington; *Geological Society of America, Bulletin*, v. 74, p. 391-406.
- Fiske, R.S. and Jackson, E.D.**
1972: Orientation and growth of Hawaiian volcanic rifts: the effect of regional structure and gravitational stresses; *Royal Society of London, Proceedings*, A329, p. 299-326.
- Fiske, R.S. and Matsuda, T.**
1964: Submarine equivalents of ash flows in the Tokiwa formation, Japan; *American Journal of Science*, v. 262, p. 76-106.
- Floyd, P.A. and Winchester, J.A.**
1975: Magma type and tectonic setting discrimination using immobile elements; *Earth and Planetary Science Letters*, v. 27, p. 211-218.
- Folinsbee, R.E., Baadsgaard, H., Cumming, G.L., and Green, D.C.**
1968: A very ancient island arc; *American Geophysical Union, Geophysical Monograph* 12, p. 441-448.

- Fontelles, M.**
1968: L'effet de socle dans le métamorphisme; Société Française de Minéralogie et de Cristallographie, Bulletin, v. 91, p. 185-206.
- Fortier, Y.O.**
1946: Yellowknife-Beaulieu region, Northwest Territories; Geological Survey of Canada, Paper 46-23.
1947: Ross Lake map-area, Northwest Territories; Geological Survey of Canada, Paper 47-16.
- Frith, R., Frith, R.A., and Doig, R.**
1977: The geochronology of the granitic rocks along the Bear-Slave Structural Province boundary, northwest Canadian Shield; *Canadian Journal of Earth Sciences*, v. 14, p. 1356-1373.
- Frith, R.A., Loveridge, W.D., and van Breemen, O.**
1986: U-Pb ages on zircon from basement granitoids of the western Slave Province, northwestern Canadian Shield; *in* Current Research, Part A, Geological Survey of Canada, Paper 86-1A, p. 113-119.
- Fyson, W.K.**
1975: Fabrics and deformation of Archean metasedimentary rocks, Ross Lake-Gordon Lake area, Slave Province, Northwest Territories; *Canadian Journal of Earth Sciences*, v. 12, p. 765-776.
1980: Fold fabrics and emplacement of a granitoid pluton, Cleft Lake, Northwest Territories; *Canadian Journal of Earth Sciences*, v. 17, p. 325-332.
1982: Complex evolution of folds and cleavages in Archean rocks, Yellowknife, N.W.T.; *Canadian Journal of Earth Sciences*, v. 19, p. 878-893.
- Fyson, W.K. and Helmstaedt, H.**
1987: Structural patterns and tectonic evolution of supracrustal domains in the Archean Slave Province, Canada; *Canadian Journal of Earth Sciences*, p. 301-315.
- Gass, I.G.**
1980: The Troodos massif: its role in the unravelling of the ophiolite problem and its significance in the understanding of constructive plate margin processes; *in* Ophiolites, ed. A. Panayiotou Proceedings of the International Ophiolite Symposium, 1979; Geological Society Department, Ministry of Agriculture and Natural Resources, Cyprus, p. 23-35.
- Gelinas, L., Brooks, C., Perrault, G., Carignan, J., Trudel, P., Grasso, F.**
1977: Chem-stratigraphic divisions within the Abitibi volcanic belt, Rouyn-Noranda District, Quebec; *in* Volcanic Regimes in Canada, ed. W.R.A. Baragar, L.C. Coleman, and J.M. Hall, J.M.; Geological Association of Canada, Special Paper Number 16, p.265-309.
- Gelinas, L., Mellinger, M., and Trudel, P.**
1982: Archean mafic metavolcanics from the Rouyn Noranda district, Abitibi greenstone belt, Quebec; I. mobility of the major elements; *Canadian Journal of Earth Sciences*, v. 19, p. 2258-2275.
- Gibbins, W.A. and Hogarth, D.D.**
1986: High magnesium or komatiitic peridotite from the Archean Hope Bay volcanic belt, Slave Province, Northwest Territories; Geological Association of Canada, Mineralogical Association of Canada, Canadian Geophysical Union, Joint Annual Meeting, Ottawa 1986, Program with Abstracts, v. 11, p. 72.
- Glassley, W.**
1974: Geochemistry and tectonics of the Crescent volcanic rocks, Olympic Peninsula, Washington; Geological Society of America, Bulletin, v. 85, p. 785-794.
- Goetz, P.Q.**
1974: Petrology and emplacement of the Detour Lake tonalite-granodiorite porphyry; University of Waterloo, unpublished B.Sc. thesis, 60 p.
- Green, D.C. and Baadsgaard, H.**
1971: Temporal evolution and petrogenesis of an Archean crustal segment at Yellowknife, N.W.T., Canada; *Journal of Petrology*, v. 12, p. 177-217.
- Green, D.C., Baadsgaard, H., and Cumming, G.L.**
1968: Geochronology of the Yellowknife area, Northwest Territories, Canada; *Canadian Journal of Earth Sciences*, v. 5, p. 725-735.
- Green, D.H. and Ringwood, A.E.**
1968: Genesis of the calc-alkaline igneous rock suite; *Contributions to Mineralogy and Petrology*, v. 18, p. 105-162.
- Grotzinger, J.P.**
1986: Evidence for early Proterozoic foredeep subsidence and sedimentation: a new interpretation of the Kilohigok basin, N.W.T., Canada; *in* Geological Association of Canada, Mineralogical Association of Canada, Canadian Geophysical Union, Joint Annual Meeting, Ottawa 1986, Program with Abstracts, v. 11, p. 76.
- Grotzinger, J.P. and Gall, G.**
1986: Preliminary investigations of Early Proterozoic Western River and Burnside River formations: evidence for foredeep origin of Kilohigok Basin, District of Mackenzie; *in* Current Research Part A, Geological Survey of Canada, Paper 86-1A, p. 95-106.
- Helmstaedt, H. and Padgham, W.A.**
1986: A new look at the stratigraphy of the Yellowknife Supergroup at Yellowknife, N.W.T. - implications of the age of gold-bearing shear zones and Archean basin evolution; *Canadian Journal of Earth Sciences*, v. 23, p. 454-475.
- Helmstaedt, H., Padgham, W.A. and Brophy, J.A.**
1986: Multiple dikes in Lower Kam Group, Yellowknife greenstone belt: evidence for Archean sea-floor spreading?; *Geology*, v. 14, p. 562-566.
- Henderson, J.B.**
1970: Stratigraphy of the Archean Yellowknife Supergroup, Yellowknife Bay-Prosperous Lake area, District of Mackenzie; Geological Survey of Canada, Paper 70-26, 12 p.
1975a: Sedimentological studies of the Yellowknife Supergroup in the Slave Structural Province; *in* Report of Activities, Part A, Geological Survey of Canada Paper 75-1A, p. 325-330.
1975b: Sedimentology of the Archean Yellowknife Supergroup at Yellowknife, District of Mackenzie; Geological Survey of Canada, Bulletin 246.
1976: Yellowknife and Hearne Lake map-areas, District of Mackenzie, Northwest Territories; Geological Survey of Canada, Open File 353.
1981: Archean basin evolution in the Slave Province, Canada; *in* Precambrian Plate Tectonics, ed. A. Kroner; Elsevier Scientific Publishing Company, Amsterdam.
1985: Geology of the Yellowknife-Hearne Lake area, District of Mackenzie: a segment across an Archean basin; Geological Survey of Canada, Memoir 414.
- Henderson, J.B. and Easton, R.M.**
1977: Archean supracrustal-basement relationships in the Keskarrah Bay map-area, Slave Province, District of Mackenzie; *in* Report of Activities, Part A, Geological Survey of Canada, Paper 77-1A, p. 217-221.
- Henderson, J.B., van Breemen, O., and Loveridge, W. D.**
1987: Some U-Pb zircon ages from Archean basement, supracrustal and intrusive rocks, Yellowknife-Hearne Lake area, District of Mackenzie; *in* Radiogenic Age and Isotope Studies: Report 1, Geological Survey of Canada Paper 87-2, p. 11-121.
- Henderson, J.B., Cecile, M.P., and Kamineni, D.C.**
1972: Yellowknife and Hearne Lake map-areas, District of Mackenzie, with emphasis on the Yellowknife Supergroup (Archean); *in* Report of Activities, Part A, Geological Survey of Canada, Paper 72-1, p. 117-119.
- Henderson, J.B., Lambert, M.B., and Peeling, G.**
1973: Yellowknife and Hearne Lake map-areas, District of Mackenzie; *in* Report of Activities, Part A, Geological Survey of Canada, Paper 73-1, p. 148-151.
- Henderson, J.B., Loveridge, W.D., and Sullivan, R.W.**
1982: A U-Pb study of zircon from granitic basement beneath the Yellowknife Supergroup, Point Lake, District of Mackenzie; *in* Rb-Sr and U-Pb Isotopic Age Studies, Report 5, *in* Current Research, Part C, Geological Survey of Canada, Paper 82-1C, p. 173-178.
- Henderson, J.F.**
1938: Beaulieu River area, Northwest Territories; Geological Survey of Canada, Paper 38-1.
1939: Beaulieu River area, Northwest Territories; Geological Survey of Canada, Paper 39-1.

- Henderson, J.F. (cont.)**
 1941: Gordon Lake south. District of Mackenzie, N.W.T.; Geological Survey of Canada, Map 645A.
 1953: On the formation of pillow lavas and breccias; Royal Society of Canada, Transactions, v. 47, Ser. III, p. 23-32.
- Henderson, J.F. and Jolliffe, A.W.**
 1941: Beaulieu River, District of Mackenzie, N.W.T.; Geological Survey of Canada, Map 581A.
- Heywood, W.W. and Davidson, A.**
 1969: Geology of Benjamin Lake map-area, District of Mackenzie; Geological Survey of Canada, Memoir 361.
- Hoffman, P.F.**
 1986: Crustal accretion in a 2.7-2.5 Ga "granite-greenstone" terrane, Slave Province, NWT: a prograding trench-arc system?; in Geological Association of Canada, Mineralogical Association of Canada, Canadian Geophysical Union, Joint Annual Meeting, Ottawa 1986, Program with Abstracts, v. II, p. 82.
- Hoffman, P.F., Culshaw, N.G., Hanmer, S.K., LeCheminant, A.N., McGrath, P.H., Tirrul, R., van Breemen, O., Bowering, S.A., and Grotzinger, P.J.**
 1986: Is the Thelon Front(NWT) a suture?; in Geological Association of Canada, Mineralogical Association of Canada, Canadian Geophysical Union, Joint Annual Meeting, Ottawa 1986, Program with Abstracts, v. II, p. 82.
- Holm, P.E.**
 1982: Non-recognition of continental tholeiites using the Ti-Y-Zr diagram; Contributions to Mineralogy and Petrology, v. 79, p. 308-310.
- Irvine, T.N. and Baragar, W.R.A.**
 1971: A guide to the chemical classification of the common volcanic rocks; Canadian Journal of Earth Sciences, v. 8, p. 523-548.
- Jakobsson, S.P.**
 1972: Chemistry and distribution pattern of recent basaltic rocks in Iceland; Lithos, v. 5, p. 365-386.
- Jensen, L.S.**
 1976: A new cation plot for classifying subalkalic volcanic rocks; Ontario Division of Mines, Miscellaneous Paper 66.
- Kamineni, D.C.**
 1975: Chemical mineralogy of some cordierite-bearing rocks near Yellowknife, Northwest Territories, Canada; Contributions to Mineralogy and Petrology, v. 53, p. 293-310.
- Kamineni, D.C., Divi, S.R., and Tella, S.**
 1979: Time relations of metamorphism and deformation in Archean metasedimentary rocks near Yellowknife, Canada; Neues Jahrbuch für Mineralogie, Monatshefte, Jahrgang 1979, p. 34-48.
- Kidd, R.G.W. and Cann, J.R.**
 1974: Chilling statistics indicate an ocean-floor spreading origin for the Troodos Complex, Cyprus; Earth and Planetary Science Letters, v. 24, p. 151-155.
- Krogh, T.E. and Gibbins, W.**
 1978: U-Pb isotopic ages of basement and supracrustal rocks in the Point Lake area of the Slave Province, Canada; in Program with Abstracts, Geological Association of Canada, Mineralogical Association of Canada, v. 3, p. 438.
- Kuno, H.**
 1968: Differentiation of basalt magmas; in Basalts, Volume 2, ed. H.H. Hess and A. Poldervaart, p. 623-688. Interscience, John Wiley and Sons, New York, 862p.
- Kushiro, I.**
 1972: Effect of water on the composition of magmas formed at high pressures; Journal of Petrology, v. 13, part 2, p. 311-334.
- Kusky, T.M.**
 1986a: Are greenstone belts in the Slave Province, N.W.T., allochthonous?; in Workshop on the Tectonic Evolution of Greenstone Belts; Lunar and Planetary Institute, Technical Report 86-10, p. 135-139, Houston.
 1986b: Ultramafic rocks and mélange units suggest an ophiolitic origin for Slave Province greenstone belts; 6th Annual Canadian Tectonics Study Group Meeting, Program with Abstracts, p. 4.
- Kusky, T.M. (cont.)**
 1986c: Thrusting between the Cameron River greenstone belt and the Sleepy Dragon metamorphic complex; Contributions to the Geology of the Northwest Territories, v. 3.
 1986d: Basement — cover relationships in the Cameron River greenstone belt, Northwest Territories, Canada; Geological Society of America, Program with Abstracts, v. 18, no. 1, p. 28.
 1987: Comment on "Multiple dykes in the lower Kam Group, Yellowknife greenstone belt: Evidence for sea-floor spreading; Geology, v. 15, no. 3, p. 280-281.
 — Archean thin skinned tectonics and allochthonous greenstone belts in the Slave Province, N.W.T.; Tectonics (in press).
- Lambert, M.B.**
 1974: Archean volcanic studies in the Slave-Bear Province; in Report of Activities, Part A, Geological Survey of Canada, Paper 74-1A, p. 177-179.
 1977: Anatomy of a greenstone belt, Slave Province, Northwest Territories; Geological Association of Canada, Special Paper 16, p. 331-340.
 1982: Synvolcanic intrusions in the Cameron River volcanic belt, District of Mackenzie; in Current Research, Part A, Geological Survey of Canada, Paper 82-1A, p. 165-167.
- Lambert, M.B. and Ernst, R.E.**
 1987: Archean mafic dyke swarm, Beaulieu River volcanic belt, Slave Province, N.W.T.; in Current Research, Part A, Geological Survey of Canada Paper 87-1A, p. 673-679.
- Lambert, M.B. and Henderson, J.B.**
 1980: A uranium-lead age of zircons from volcanics and sediments of the Back River volcanic complex, Eastern Slave Province, District of Mackenzie; in Rubidium-strontium and uranium-lead isotopic age studies, Report 3; in Current Research, Part C, Geological Survey of Canada, Paper 80-1C, p. 239-242.
- Lambert, M.B. and van Staal, C.R.**
 1987: Archean granite-greenstone boundary relationships in the Beaulieu River volcanic belt; in Current Research, Part A, Geological Survey of Canada Paper 87-1A, p. 605-618.
- Lambert, R. St. John**
 1974: Chemical petrology of a suite of calc-alkaline lavas from Mount Ararat, Turkey; Journal of Geology, v. 82, p. 419-438.
- Liou, J.G., Kuniyoshi, S., and Ito, K.**
 1974: Experimental studies of the phase relations between greenschist and amphibolite in a basaltic system; American Journal of Science, v. 274, p. 613-632.
- Lonsdale, P. and Spiess, F.N.**
 1979: A pair of young cratered volcanoes on the East Pacific Rise; Journal of Geology, v. 87, p. 157-173.
- Lord, C.S.**
 1951: Mineral industry of District of Mackenzie, Northwest Territories; Geological Survey of Canada, Memoir 261.
- Ludden, J., Gelinas, L., and Trudel, P.**
 1982: Archean metavolcanics from the Rouyn-Noranda District, Abitibi greenstone belt, Quebec: 2. mobility of trace elements and petrogenetic constraints; Canadian Journal of Earth Sciences, v. 19, p. 2276-2287.
- Marsh, D.M.**
 1982: On the mechanics of igneous diapirism, stoping, and zone melting; American Journal of Science, v. 282, p. 808-855.
 1984: Mechanics and energetics of magma formation and ascent; in Explosive Volcanism: Studies in Geophysics, National Academy Press, Washington, D.C., p. 67-83.
- Maruyama, S., Suzuki, K., and Liou, J.G.**
 1983: Greenschist-amphibolite transition equilibria at low pressure; Journal of Petrology, v. 24, p. 583-604.
- MacGeehan, P.J.**
 1978: The geochemistry of altered volcanic rocks at Matagami, Québec: a geothermal model for massive sulphide genesis; Canadian Journal of Earth Sciences, v. 15, p. 551-570.
- McBirney, A.R.**
 1963: Factors governing the nature of submarine volcanism; Bulletin Volcanologique, v. 26, p. 455-469.

- McGlynn, J.C.**
1977: Geology of Bear-Slave Structural Provinces, District of Mackenzie; Geological Survey of Canada, Open File 445.
- McGlynn, J.C. and Henderson, J.B.**
1970: Archean volcanism and sedimentation in the Slave Structural Province; in *Symposium on Basins and Geosynclines of the Canadian Shield*, ed. A.J. Baer; Geological Survey of Canada, Paper 70-40, p. 31-44.
1972: The Slave Province; in *Variations in Tectonic Styles in Canada*, ed. R.A. Price and R.J.W. Douglas; Geological Association of Canada, Special Paper 11, p. 506-526.
- McGlynn, J.C. and Irvine, E.**
1975: Paleomagnetism of early Aphebian diabase dykes from the Slave Province, Canada; *Tectonophysics*, v. 26, p. 23-38.
- McGrath, P.H., Henderson, J.B., and Lindia, F.M.**
1983: Interpretation of a gravity profile over a contact zone between an Archean granodiorite and the Yellowknife Supergroup using an interactive computer program with partial automatic optimization; in *Current Research, Part B*, Geological Survey of Canada, Paper 83-1B, p. 189-194.
- Miyashiro, A.**
1968: Metamorphism of mafic rocks; in *Basalts*, ed. H.H. Hess and A. Poldervaart; New York, Interscience, v. 2, p. 799-834.
- Moody, J.B., Meyer D., and Jenkins, J.E.**
1983: Experimental characterization of the greenschist/amphibolite boundary in mafic rocks; *American Journal of Science*, v. 283, p. 48-92.
- Moore, J.G.**
1965: Petrology of deep-sea basalt near Hawaii; *American Journal of Science*, v. 263, p. 40-52.
- Nikic, Z., Baadsgaard, H., Folinsbee, R.E., Krupicka, J., Leech, A.P., and Sasaki, A.**
1980: Boulders from the basement, the trace of an ancient crust?; in *Selected studies of Archean Gneisses and Lower Proterozoic Rocks, Southern Canadian Shield*, ed. G.B. Morey and G.N. Hanson; Geological Society of America, Special Paper 182, p. 169-175.
- Osborn, E.F.**
1959: Role of oxygen pressure in the crystallization and differentiation of basaltic magma; *American Journal of Science*, v. 257, p. 609-647.
- Padgham, W.A.**
1985: Observations and speculations on supracrustal successions in the Slave Structural Province; in *Evolution of Archean Supracrustal Sequences*, ed. L.D. Ayres, P.C. Thurston, K.D. Card, and W. Weber; Geological Association of Canada, Special Paper 28, 1985, p. 133-151.
- Pearce, J.A. and Cann, J.R.**
1971: Ophiolite origin investigated by discriminant analysis using Ti, Zr and Y; *Earth and Planetary Science Letters*, v. 12, p. 339-349.
1973: Tectonic setting of basic volcanic rocks determined using trace element analysis; *Earth and Planetary Science Letters*, v. 19, p. 290-300.
- Pearce, J.A. and Norry, M.J.**
1979: Petrogenetic implications of Ti, Zr, Y, and Nb variations in volcanic rocks; *Contributions to Mineralogy and Petrology*, v. 69, p. 33-47.
- Pearce, T.H., Gorman, B.E., and Birkett, T.C.**
1975: The TiO_2 -K₂O-P₂O₅ diagram: a method of discriminating between oceanic and non-oceanic basalts; *Earth and Planetary Science Letters*, v. 24, p. 419-426.
- Platt, J.P.**
1980: Archean greenstone belts: a structural test of tectonic hypotheses; *Tectonophysics*, v. 65, p. 127-150.
- Ramsay, C.R. and Kameneni, C.**
1977: Petrology and evolution of an Archean metamorphic aureole in the Slave craton, Canada; *Journal of Petrology*, v. 18, p. 460-486.
- Ramsay, J.G.**
1967: *Folding and Fracturing of Rocks*; McGraw Hill, 568p.
- Ramsay, J.G. and Huber, M.I.**
1987: *The Techniques of Modern Structural Geology, Volume 2: Folds and Fractures*; Academic Press, Toronto, p. 309-700.
- Rudman, A.J. and Epp, D.**
1983: Conduction models of the temperature distribution in the East Rift Zone of Kilauea Volcano; *Journal of Volcanology and Geothermal Research*, v. 16, p. 189-203.
- Sakuyama, M.**
1983: Petrology of arc volcanic rocks and their origin by mantle diapirs; *Journal of Volcanology and Geothermal Research*, v. 18, p. 297-320.
- Schärer, U. and Allègre, C.J.**
1982: Investigation of the Archean crust by single-grain dating of detrital zircon: a greywacke of the Slave Province, Canada; *Canadian Journal of Earth Sciences*, v. 19, p. 1910-1918.
- Schau, M. and Henderson, J.B.**
1983: Archean chemical weathering at three localities on the Canadian Shield; *Precambrian Research*, v. 20, p. 189-224.
- Schiener, E.L.**
1974: Syndepositional small-scale intrusions in Ordovician pyroclastics, Co. Waterford, Ireland; *Geological Society of London, Journal*, v. 130, p. 157-161.
- Shaw, H.R.**
1973: Mantle convection and volcanic periodicity in the Pacific: evidence from Hawaii; *Geological Society of America, Bulletin*, v. 84, p. 1505-1526.
- Sibson, R.H.**
1977: Fault rocks and fault mechanisms; *Geological Society of London, Journal*, v. 133, p. 191-213.
- Sigurdsson, H.**
1985: Dyke injection in Iceland: a review; in *Mafic Dyke Swarms*, ed. H.C. Halls and W.F. Fahrig; Geological Association of Canada, Special Paper 34, p. 55-64.
- Sigurdsson, H. and Sparks, R.S.J.**
1978: Rifting episode in north Iceland in 1874-1875 and the eruptions of Askja and Sveinagja; *Bulletin Volcanologique*, v. 41, p. 149-167.
- Sigurdsson, H., Sparks, R.S.J., Carey, S.N., and Huang, T.C.**
1980: Volcanogenic sedimentation in the Lesser Antilles Arc; *Journal of Geology*, v. 88, p. 523-540.
- Simpson, C. and Schmid, S.M.**
1983: An evaluation of criteria to deduce the sense of movement in sheared rocks; *Geological Society of America, Bulletin*, v. 94, p. 1281-1288.
- Smith, R.E. and Smith, S.E.**
1976: Comments on the use of Ti, Zr, Y, Sr, K, P, and Nb in classification of basaltic magmas; *Earth and Planetary Science Letters*, v. 32, p. 114-120.
- Smith, R.L.**
1960: Zones and zonal variations in welded ash flows; *U.S. Geological Survey, Professional Paper 354F*, p. 149-159.
- Steiger, R.H. and Jaeger, E.**
1977: Subcommission on geochronology: convention on the use of decay constants in geo- and cosmo-chronology; *Earth and Planetary Sciences Letters*, v. 36, p. 359-362.
- Stockwell, C.H.**
1933: Great Slave Lake-Coppermine River area, Northwest Territories; *Geological Survey of Canada, Summary Report, 1932, Pt. C*, p. 37-63.
1962: Interpretation of potassium-argon isotopic age GSC 60-49; in *Age Determinations by the Geological Survey of Canada, Report 2, Isotopic Ages*, ed. J.A. Lowdon; Geological Survey of Canada, Paper 61-17, p. 29.
- Stockwell, C.H., McGlynn, J.C., Emslie, R.F., Sanford, B.V., Norris, A.W., Donaldson, J.A., Fahrig, W.F., and Currie, K.L.**
1968: Geology of the Canadian Shield; in *Geology and Economic Minerals of Canada*, ed. R.J.W. Douglas; Geological Survey of Canada, Economic Geology Report No. 1, p. 44-150.

St-Onge, M.R., King, J.E., and Lalonde, A.E.

- 1984: Deformation and metamorphism of the Coronation Supergroup and its basement in the Hepburn metamorphic-plutonic zone of Wopmay Orogen: Redrock Lake and the eastern portion of Calder River map areas, District of Mackenzie; in *Current Research, Part A*, Geological Survey of Canada, Paper 84-1A, p. 171-180.

Streckeisen, A.L.

- 1967: Classification and nomenclature of igneous rocks; *Neues Jahrbuch für Geologie und Paläontologie: Abhandlungen*, Stuttgart, v. 107, p. 144-240.

Swanson, D.A.

- 1972: Magma supply rate at Kilauea Volcano, 1952-1971; *Science*, v. 175, p. 169-170.

Swanson, D.A., Duffield, W.A., and Fiske, R.S.

- 1976: Displacement of the south flank of Kilauea Volcano: the result of forceful intrusion of magma into the rift zones; *U.S. Geological Survey, Professional Paper 963*, 39p.

Tarney, J., Dalziel, I.W.D., and DeWit, M.J.

- 1976: Marginal basin 'Rocas Verdes' complex from S. Chile: a model for Archean greenstone belt formation; in *The Early History of the Earth*, ed. B. F. Windley; p. 131-146.

Thompson, P.H.

- 1978: Archean regional metamorphism in the Slave Structural Province — a new perspective on some old rocks; in *Metamorphism in the Canadian Shield*, ed. J.A. Fraser and W.W. Heywood; Geological Survey of Canada, Paper 78-10, p. 85-102.

Thorarinsson, S.

- 1967: Some problems of volcanism in Iceland; *Geologische Rundschau*, v. 57, p. 1-20.

Thorpe, R.I.

- 1971: Comments on rock ages in the Yellowknife area, District of Mackenzie; in *Report of Activities, Part B*, Geological Survey of Canada, Paper 71-1B, p. 76-79.

- 1982: Lead isotope evidence regarding Archean and Proterozoic metallogeny in Canada; *Revista Brasileira de Geociencias*, v. 12, p. 510-521.

Tobisch, O.T., Saleeby, J.B., and Fiske, R.S.

- 1986: Structural history of continental volcanic arc rocks, Eastern Sierra Nevada, California: a case for extensional tectonics; *Tectonics*, v. 5, p. 65-94.

Vejnar, Z.

- 1977: The relationship between the metamorphic grade and composition of silicates in the west Bohemian greenschists and amphibolites; *Krystalinikum*, v. 13, p. 129-158.

Vogt, P.R.

- 1976: Plumes, subaxial pipe flow, and topography along the mid-oceanic ridge; *Earth and Planetary Sciences Letters*, v. 29, p. 309-325.

Walker, G.P.L.

- 1986: Koolau Dike Complex, Oahu: intensity and origin of a sheeted-dike complex high in a Hawaiian volcanic edifice; *Geology*, v. 14, p. 310-313.

- 1987: The dyke complex of Koolau volcano, Oahu: internal structure of a Hawaiian rift zone; in *Volcanism in Hawaii*, ed. R.W. Decker, T.L. Wright and P.H. Stauffer; U.S. Geological Survey, Professional Paper 1350, p. 961-993.

White, S.H., Burrows, S.E., Carreras, J., Shaw, N.D., and Humphreys, F.J.

- 1980: On mylonites in ductile shear zones; *Journal of Structural Geology*, v. 2, p. 175-187.

Winchester, J.A. and Floyd, P.A.

- 1976: Geochemical magma type discrimination: applied to altered and metamorphosed basic igneous rocks; *Earth and Planetary Science Letters*, v. 28, p. 459-469.

- 1977: Geochemical discrimination of different magma series and their differentiation products using immobile elements; *Chemical Geology*, v. 20, p. 325-343.

- 1984: The geochemistry of the Ben Hope sill suite, northern Scotland, U.K.; *Chemical Geology*, v. 43, p. 49-75.

Zeck, H.P. and Morthorst, J.R.

- 1982: Continental tholeiites in the Ti-Zr-Y discrimination diagram; *Neues Jahrbuch für Mineralogie, Monatshefte*, v. 5, p. 193-200.

Ziegler, P.A.

- 1982: Faulting and graben formation in western and central Europe; *Royal Society of London, Philosophical Transactions*, v. 305, p. 113-143.

APPENDIX I

Mode and texture of granitic rocks near the
Cameron River and Beaulieu River Volcanic Belts.

Specimen no.	Texture	Quartz	Plagioclase	Microcline	Biotite	Muscovite	Chlorite (hornblende)	Opaque	Accessory
1,2 etc. code	f	g	n	pt	po preferred orientation			mt	a
numbers in Fig. 5	m	texture	o	g				ht	p
*mode from stained slab	ph		my	pt				py	e
+ mode from Goets (1974)	al		a	g					c
			sa						s
									x
									z
									g
									tr
									amount
DEFEAT PLUTONIC SUITE AD									
1	m, ph, bi, hy	26.8	62.2 (An ₁₀) _{my,aa}	6.1 pt	2.9	tr	0.2	0.5-mt	1.0+tr-a, z
2	m, ph, bi, hy	28.2	56.2 (An ₁₀) _{my,aa}	11.3 pt	1.9			1.4-py	0.5+tr-a, z
3	m, ph, bi, hy	34.9	39.0 (An ₁₀) _{my,aa}	21.4 pt	2.7 po	1.1 po	0.4	tr-mg	0.5+tr-p, c, a, g
4	m, bi	41.3	42.0 (An ₁₀) _{my}	14.6 pt	1.9 po	0.8 po	tr	0.5-mt	tr-a, z
5	c, ph, bi, Mi	45.0	41.4 (An ₁₀) _{my,aa}	4.0	8.2 po		tr(0.7)	tr-py	0.5+tr-p, a, g
6	m, bi	20.2	53.8 (An ₁₀) _{my,aa}	13.6 pt	11.8	0.1	x	tr-py	0.5+tr-a, z
MEANDER LAKE PLUTONIC SUITE AML									
7	m, bi	37.9	44.0 (An ₁₀) _{my,aa}	16.2 pt	1.0	0.8	tr	tr-p, c, g	tr-p, c, g
8*	m, bi	30.6	46.4	11.7	2.7			tr-py	0.5+tr-a, z
9	m, ph, bi, hy	19.9	38.2 (An ₂₁) _{aa}	11.7	5.3	0.4	tr(32.1)	tr-py	0.5+tr-a, z
10*	m, bi	28.5	41.0	18.8	1.5			tr-py	1.3+tr-o
11	m, bi	39.5	31.0 (An ₁₃) _{aa}	23.7 pt	1.1	2.7	0.6	tr-py	tr-p, c, g
AMACHER GRANITE AAM									
12	m, bi, hy	37.8	14.3 (An ₁₀) _{my}	47.2 pt	1.2	0.2	0.4	tr-mt	tr-p, c, g
13	m, ph, bi, hy	39.5	37.7 (An ₁₀) _{my,aa}	14.5 pt	3.4		4.4	tr-py	0.5+tr-a, z
14*	c, bi	45.8	24.8	25.6	3.7			tr-py	0.5+tr-a, z
15	c, bi	44.2	15.9 (An ₁₃) _{aa}	34.5 pt	2.1	1.2	1.2	tr-py	0.5+tr-a, z
16	c, bi	48.1	18.2 (An ₁₃) _{aa}	16.2 pt, g		12.5 aa	2.0	2.8	tr-a, z
DETOUT GRANODIORITE ADT									
17+	c, ph, bi, hy	49.05	35.54	9.24	4.15				
18+	c, ph, bi, hy	46.72	37.91	5.54	7.30				
19+	c, ph, bi, hy	39.60	46.64	2.21	8.36				
20+	c, ph, bi, hy	43.83	45.58	2.79	5.36				
REDOUT GRANITE AR									
21	c, ph, bi, hy	34.8	35.7 (An ₁₀) _{my,aa}	16.8 pt, g	4.0 po	6.9 po	tr	tr-mt	tr-a, z
22	c, ph, bi, hy	37.3	18.8	52.2	1.4	0.7		tr-mt	tr-p, c, a, g
23	m, ph, al	44.1	28.2 (An ₁₃) _{aa}	15.3 aa	3.2 po	8.7 po	0.3	tr-mt	tr-p, c, a, g
24*	c, bi	40.2	32.4	24.1	3.3				
25*	m, ph, bi, hy	41.6	42.5 aa	15.4	0.9				
26*	m, bi, ph, Mi	35.1	43.8	18.9	2.2 po		tr(tr)		
PROSPEROUS GRANITE AP									
27*	m, bi, ph, Mi	39.2	41.5	11.8	7.5 po				
28*	m, bi, ph, Mi	32.6	49.4	16.4	1.5				
29	m, ph, bi, Mi	29.4	32.8 (An ₁₀) _{aa}	28.6	tr	9.2	tr	tr-p, g	tr-p, g
30	m, ph, bi, hy	33.6	36.2 (An ₁₀) _{aa}	14.7 pt	0.1	15.3 po	0.1	tr-py	tr-p
31*	m, bi	47.2	29.9	14.0	2.0	11.7 po	3.0	tr-py	tr-p
32*	m, bi	32.9	48.2	12.5	6.4			0.6-mt	z, g
33	m, ph, bi, Mi, hy	47.20	25.60 (An ₁₀) _{my}	18.01	3.24	5.59	0.3	tr	z, g
34	m, ph, bi, hy	27.76	51.60 (An ₂₀) _{my}	11.98	5.46	3.20	tr	tr	z, g
35	m, ph, bi, Mi, hy	31.78	30.30 (An ₁₃) _{my}	32.37 pt	3.5	2.05	tr	tr	tr-p, z, a

APPENDIX 2

Chemical analyses, specific gravities, normative compositions, and SiO₂ variation diagrams of rocks from the Cameron River-Beaulieu River volcanic belts.

- A. Major oxides and specific gravity
- B. Minor and trace elements
- C. Normative compositions
- D. SiO₂ variation diagrams

Dot (.) beside sample numbers (IDENT) indicates least altered samples.

Symbols in parentheses following formation names are symbols used in all chemical plots.

Mafic dykes are listed with the Cameron River Basalt. These dykes are from all belts. Letters following sample numbers indicate the belt where dykes occur: C — Cameron River subarea; T — Tumpline Lake subarea; S — Sunset Lake subarea.

Specimen 385 (Alice Formation, Dacite Member) is a felsic dyke that may not be related to this formation.

AV is the average of analyses listed for each formation rounded to the third decimal for minor elements.

APPENDIX 2A

Major oxides and specific gravity

CAMERON RIVER SUBAREA

Rhyolite Ar X

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
.379	.0170	.0390	.0034	.0210	.0092	.0020	.0030	.0010	.0005	.0011	70.0	24.0	16.0	3.3	.0510
.414	.0050	.0050	.0000	.0100	.0000	.0005	.0003	.0022	.0005	.0000	54.0	10.0	21.0	5.0	.0250
.B1	.0160	.0120	.0160	.0000	.0270	.0120	.0088		.0010	.0100	22.0	1.8	13.0	1.7	.1300
AV	.0127	.0187	.0065	.0103	.0121	.0058	.0040	.0016	.0007	.0037	48.7	11.9	16.7	3.3	.0687

Webb Lake Andesite AW ◇

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
.377	.0140	.0080	.0140	.0230	.0170	.0090	.0047	.0025	.0025	.0023	120.0	2.7	15.0	2.3	.0960
.415	.0150	.0020	.0180	.0180	.0220	.0110	.0039	.0010	.0032	.0024	82.0	2.2	16.0	1.2	.0250
.B3	.0130	.0047	.0210	.0000	.0260	.0210	.0078		.0120	.0130	110.0	6.2	21.0	1.3	.0710
.B4	.0120	.0110	.0230	.0000	.0240	.0120	.0063		.0000	.0130	80.0	5.2	17.0	0.0	.0900
.B5	.0190	.0460	.0000	.0200	.0000	.0000	.0000		.0000	.0000	75.0	6.6	20.0	0.0	.1600
.B10	.0077	.0200	.0130	.0110	.0110	.0085	.0033		.0000	.0000	39.0	6.2	19.0	1.5	.1500
.B11	.0220	.0380	.0190	.0130	.0180	.0170	.0074		.0000	.0000	94.0	5.9	13.0	1.9	.1300
.B12	.0240	.0260	.0190	.0110	.0180	.0180	.0078		.0000	.0050	78.0	5.8	10.0	1.0	.1200
AV	.0158	.0195	.0159	.0120	.0170	.0121	.0052	.0018	.0022	.0051	84.8	5.9	16.4	1.5	.1053

Cameron River Basalt AC ○

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
.380	.0110	.0090	.0240	.0090	.0370	.0130	.0140	.0010	.0047	.0043	85.0	.3	14.0	1.4	.0560
.386	.0110	.0070	.0210	.0110	.0340	.0120	.0018	.0024	.0042	.0039	52.0	2.1	13.0	.6	.0250
.387	.0020	.0040	.1300	.0080	.0230	.0240	.0000	.0010	.0061	.0038	110.0	.3	15.0	.8	.0250
.388	.0130	.0040	.0590	.0130	.0390	.0170	.0030	.0021	.0051	.0047	77.0	.3	16.0	1.7	.0250
.389	.0110	.0050	.0840	.0140	.0430	.0220	.0017	.0022	.0048	.0056	92.0	.3	16.0	3.0	.0250
.390	.0110	.0040	.0045	.0200	.0470	.0037	.0083	.0036	.0030	.0037	92.0	.3	15.0	3.3	.0250
.391	.0170	.0040	.0055	.0220	.0530	.0068	.0066	.0041	.0044	.0040	87.0	.3	13.0	1.4	.0250
.392	.0170	.0040	.0360	.0100	.0360	.0110	.0100	.0027	.0047	.0049	56.0	.3	13.0	.8	.0250
.400	.0090	.0060	.0160	.0100	.0450	.0170	.0011	.0023	.0044	.0042	130.0	3.6	14.0	2.5	.0250
.401	.0100	.0150	.0190	.0100	.0380	.0110	.0089	.0023	.0052	.0041	120.0	.3	18.0	1.1	.0250
.402C	.0130	.0040	.0170	.0100	.0390	.0110	.0120	.0024	.0043	.0037	82.0	.3	16.0	1.3	.0250
.402X	.0130	.0040	.0170	.0100	.0390	.0110	.0120	.0024	.0043	.0037	82.0	.3	16.0	1.3	.0250
.403	.0110	.0040	.0190	.0130	.0400	.0110	.0051	.0021	.0046	.0036	73.0	.3	14.0	1.0	.0250
.404	.0200	.0200	.0190	.0120	.0410	.0120	.0027	.0010	.0038	.0040	76.0	.3	14.0	1.0	.0250
.405	.0120	.0050	.0190	.0120	.0400	.0110	.0051	.0021	.0046	.0036	73.0	.3	17.0	1.2	.0250
.406	.0130	.0030	.0120	.0100	.0390	.0097	.0150	.0010	.0050	.0043	120.0	.3	16.0	1.0	.0650
.407	.0130	.0040	.0110	.0110	.0410	.0110	.0140	.0020	.0057	.0048	87.0	.3	15.0	1.2	.0250
.424	.0120	.0030	.0200	.0100	.0390	.0100	.0036	.0027	.0043	.0042	150.0	2.2	20.0	.9	.0250
.425	.0130	.0040	.0230	.0100	.0400	.0120	.0097	.0027	.0051	.0046	140.0	2.1	18.0	1.0	.0530
.434X	.0140	.0040	.0180	.0110	.0400	.0096	.0110	.0021	.0041	.0042	98.0	.3	16.0	1.0	.2500
.B6	.0210	.0510	.0000	.0100	.0000	.0000	.0000		.0000	.0000	90.0	1.8	20.0	1.1	.0900
.B7	.0150	.0110	.0110	.0110	.0260	.0130	.0060		.0057	.0083	73.0	5.0	22.0	1.3	.0680
.B8	.0200	.0320	.0140	.0140	.0250	.0120	.0052		.0010	.0020	120.0	2.3	14.0	1.1	.1100
.B14	.0057	.0100	.0380	.0000	.0220	.0220	.0160		.0083	.0050	54.0	2.3	17.0	.8	.0820
AV	.0128	.0092	.0270	.0103	.0359	.0122	.0075	.0022	.0044	.0041	94.0	1.0	15.9	1.3	.0489

Mafic Dykes Aa X

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
.399	.0100	.0040	.0270	.0070	.0270	.0210	.0180	.0010	.0057	.0032	75.0	.3	14.0	1.0	.0740
.432	.0410	.0280	.0053	.0330	.0530	.0098	.0340	.0039	.0051	.0026	110.0	3.3	21.0	2.6	.1200
.566					.0290	.0000	.0074	.0025	.0039	.0023	0.0	.3	13.0	1.7	.0250
.B2	.0110	.0130	.0210	.0000	.0270	.0015	.0089		.0000	.0140	130.0	6.7	24.0	1.2	.1100
.B9	.0100	.0042	.0140	.0056	.0200	.0190	.0210		.0000	.0050	120.0	2.3	14.0	1.0	.1100
.B13	.0140	.0057	.0250	.0065	.0330	.0200	.0066		.0055	.0050	56.0	2.0	7.8	.2	.0900
AV	.0172	.0110	.0185	.0104	.0315	.0131	.0160	.0025	.0034	.0054	98.2	2.5	15.6	1.3	.0882

APPENDIX 2A (cont.)

TUMPLINE LAKE SUBAREA

Sharrie Rhyolite **ASH +**

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
559					.0000		.0000	.0048	.0000	.0000		11.0	20.0	3.9	.0250
560					.0000		.0006	.0041	.0000	.0000		22.0	12.0	3.4	.0250
AV					.0000		.0003	.0045	.0000	.0000		16.5	16.0	3.7	.0250

Turnback Rhyolite **ATB Δ**

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
378	.0100	.0290	.0005	.0200	.0000	.0005	.0091	.0024	.0005	.0019	48.0	14.0	11.0	6.5	.0250
393	.0460	.0400	.0005	.0270	.0010	.0000	.0003	.0060	.0000	.0003	92.0	18.0	16.0	9.4	.0250
394	.0050	.0300	.0005	.0390	.0010	.0000	.0013	.0047	.0005	.0014	88.0	8.0	21.0	3.7	.0250
539					.0000		.0025	.0059	.0000	.0025		14.0	21.0	5.1	.0250
561					.0000		.0000	.0052	.0000	.0000		5.7	18.0	3.4	.0250
563					.0000		.0009	.0040	.0000	.0000		18.0	13.0	3.1	.0250
564					.0000		.0015	.0026	.0000	.0000		18.0	20.0	3.8	.0750
565					.0000		.0000	.0038	.0000	.0011		5.0	12.0	2.3	.0250
572					.0000		.0000	.0029	.0000	.0008		2.3	9.7	1.2	.0250
AV	.0203	.0330	.0005	.0287	.0002	.0002	.0017	.0042	.0001	.0009	76.0	11.4	15.7	4.3	.0306

Tumpline Basalt - Andesite Member **ATM-a ●**

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
540					.0200		.0073	.0023	.0017	.0021		5.7	12.0	2.1	.0250
567					.0020		.0016	.0036	.0005	.0030		24.0	18.0	2.7	.0700
568					.0027		.0025	.0041	.0011	.0028		9.0	25.0	2.9	.0250
569					.0010		.0008	.0038	.0005	.0024		8.2	16.0	2.6	.0250
AV					.0064		.0031	.0035	.0010	.0026		11.7	17.8	2.6	.0363

Tumpline Basalt **ATM □**

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
395	.0180	.0170	.0130	.0180	.0400	.0081	.0009	.0026	.0040	.0032	100.0	2.7	12.0	1.4	.0250
396	.0160	.0160	.0140	.0160	.0350	.0096	.0180	.0020	.0028	.0033	110.0	3.1	13.0	3.5	.1100
AV	.0170	.0165	.0135	.0170	.0375	.0089	.0095	.0023	.0034	.0033	105.0	2.9	12.5	2.5	.0675

SUNSET LAKE SUBAREA

Rhyolite **AR ■**

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
384	.0030	.0710	.0000	.0130	.0000	.0005	.0046	.0026	.0000	.0000	23.0	11.0	19.0	2.5	.0250
412	.0060	.0140	.0000	.0230	.0000	.0005	.0003	.0010	.0000	.0000	5.0	25.0	9.1	1.7	.0250
413C	.0070	.0200	.0000	.0300	.0000	.0005	.0003	.0036	.0005	.0003	46.0	5.9	22.0	4.4	.0250
413X	.0070	.0200	.0000	.0300	.0000	.0005	.0003	.0036	.0005	.0003	46.0	5.9	22.0	4.4	.0250
433	.0060	.0440	.0005	.0280	.0010	.0005	.0003	.0038	.0000	.0003	52.0	6.2	23.0	4.2	.2500
562							.0003	.0057	.0000	.0008		30.0	19.0	4.8	.0250
AV	.0058	.0338	.0001	.0248	.0002	.0005	.0010	.0034	.0002	.0003	34.4	14.0	19.0	3.7	.0625

Alice Formation - Dacite Member **AAL-d ◆**

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
385	.0020	.0520	.0000	.0330	.0000	.0005	.0070	.0055	.0000	.0061	77.0	17.0	22.0	2.6	.0250
398	.0130	.0400	.0022	.0320	.0500	.0032	.0071	.0060	.0046	.0038	120.0	11.0	11.0	2.2	.0250
534					.0160		.0025	.0010	.0013	.0020		3.7	20.0	2.5	.0250
535					.0190		.0046	.0010	.0005	.0017		3.4	17.0	2.0	.0530
542					.0170		.0018	.0010	.0012	.0018		7.3	17.0	2.3	.0250
543					.0098		.0013	.0010	.0005	.0013		6.8	22.0	2.1	.0530
547					.0190		.0028	.0023	.0012	.0021		7.4	21.0	1.4	.0730
548					.0270		.0096	.0036	.0025	.0033		1.6	16.0	1.0	.0250
549					.0190		.0038	.0010	.0005	.0024		9.4	17.0	1.2	.0250
552					.0110		.0039	.0010	.0005	.0013		4.6	19.0	1.1	.0250
556					.0250		.0070	.0026	.0023	.0027		2.8	19.0	1.9	.0780
AV	.0075	.0460	.0011	.0325	.0086	.0019	.0036	.0000	.0000	.0011		2.7	19.0	1.3	.0880
					.0185		.0046	.0022	.0013	.0025	98.5	6.5	18.3	1.8	.0430

Alice Formation - Andesite Member **AAL-a +**

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
381	.0250	.0130	.0017	.0190	.0200	.0047	.0037	.0010	.0021	.0020	98.0	3.2	14.0	1.4	.0250
382	.0230	.0100	.0019	.0200	.0210	.0059	.0047	.0010	.0025	.0021	77.0	4.3	22.0	1.3	.0250
383	.0180	.0280	.0180	.0210	.0240	.0100	.0015	.0025	.0027	.0028	70.0	4.2	20.0	1.3	.0250
408	.0050	.0310	.0170	.0180	.0260	.0130	.0003	.0010	.0023	.0023	98.0	1.5	14.0	3.4	.0250
409	.0040	.0720	.0190	.0140	.0310	.0150	.0021	.0010	.0035	.0024	110.0	4.0	11.0	3.3	.0250
410	.0100	.0190	.0066	.0170	.0190	.0041	.0093	.0010	.0015	.0016	200.0	7.4	17.0	4.0	.0980
411	.0140	.0260	.0084	.0180	.0170	.0035	.0057	.0010	.0014	.0017	96.0	6.8	18.0	1.9	.0580
430	.0130	.0170	.0031	.0230	.0230	.0040	.0003	.0026	.0017	.0024	59.0	2.4	21.0	8.8	.0250
431	.0130	.0220	.0034	.0210	.0250	.0050	.0003	.0026	.0019	.0022	61.0	1.6	18.0	15.0	.0250
537					.0180		.0076	.0010	.0016	.0020		12.0	14.0	1.6	.0850
538					.0240		.0047	.0010	.0026	.0025		2.7	12.0	2.2	.0250
541					.0160		.0017	.0038	.0015	.0018		4.6	16.0	2.4	.0250
545					.0300		.0031	.0027	.0015	.0022		12.0	21.0	1.2	.0250
546					.0220		.0026	.0010	.0023	.0023		4.0	18.0	1.4	.0250
550					.0130		.0044	.0010	.0005	.0019		9.7	17.0	1.4	.0600
551					.0130		.0045	.0023	.0005	.0017		7.5	18.0	2.2	.0510
553					.0230		.0000	.0010	.0029	.0031		2.0	20.0	2.1	.1300
554					.0190		.0032	.0010	.0018	.0026		1.9	18.0	8	.0250
555					.0120		.0052	.0031	.0015	.0019		4.9	20.0	2.1	.1300
570					.0230		.0029	.0010	.0005	.0028		4.1	18.0	1.5	.0250
571					.0210		.0048	.0010	.0021	.0029		2.5	11.0	1.3	.0250
AV	.0139	.0264	.0088	.0190	.0210	.0072	.0035	.0016	.0019	.0023	96.6	4.9	17.1	3.1	.0412

APPENDIX 2A (cont.)

Sunset Lake Basalt ASU ▲

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
373C	.0070	.0020	.0170	.0100	.0320	.0067	.0140	.0010	.0037	.0037	93.0	.3	16.0	1.4	.0510
373X	.0070	.0020	.0170	.0100	.0320	.0067	.0140	.0010	.0037	.0037	93.0	.3	16.0	1.4	.0510
374	.0080	.0290	.0150	.0080	.0280	.0052	.0100	.0010	.0035	.0037	120.0	.3	16.0	1.5	.0600
375	.0130	.0060	.0260	.0090	.0310	.0160	.0065	.0010	.0045	.0040	130.0	1.6	20.0	1.7	.0510
376	.0130	.0060	.0270	.0090	.0310	.0140	.0130	.0010	.0045	.0040	140.0	3.9	19.0	1.5	.1100
416	.0070	.0130	.0320	.0070	.0310	.0220	.0110	.0010	.0050	.0040	92.0	.3	10.0	.6	.0250
417	.0100	.0040	.0380	.0090	.0340	.0150	.0091	.0010	.0055	.0043	110.0	.3	23.0	.9	.0250
418C	.0130	.0020	.0380	.0100	.0350	.0180	.0100	.0010	.0056	.0036	96.0	.3	21.0	.8	.0250
418X	.0130	.0020	.0380	.0100	.0350	.0180	.0100	.0010	.0056	.0036	96.0	.3	21.0	.8	.0250
419	.0100	.0360	.0600	.0100	.0360	.0200	.0120	.0010	.0061	.0036	100.0	.3	14.0	.7	.0250
420C	.0130	.0330	.0430	.0070	.0240	.0190	.0094	.0010	.0055	.0028	110.0	.3	26.0	1.0	.0250
420X	.0130	.0330	.0430	.0070	.0240	.0190	.0094	.0010	.0055	.0028	110.0	.3	26.0	1.0	.0250
421	.0140	.0190	.0480	.0080	.0280	.0190	.0080	.0010	.0052	.0034	81.0	.3	14.0	.7	.0250
422	.0120	.0520	.0370	.0090	.0350	.0130	.0120	.0020	.0047	.0042	100.0	.3	19.0	.7	.0250
423	.0060	.0060	.0320	.0100	.0360	.0120	.0045	.0024	.0041	.0042	150.0	.3	28.0	1.7	.0250
426	.0090	.0170	.0000	.0290	.0110	.0005	.0003	.0046	.0021	.0030	78.0	.3	21.0	5.4	.0250
427	.0090	.0140	.0000	.0300	.0110	.0005	.0003	.0046	.0021	.0030	78.0	.3	21.0	5.4	.0250
428	.0180	.0180	.0120	.0150	.0250	.0130	.0240	.0010	.0047	.0024	96.0	4.1	30.0	2.9	.0250
429	.0130	.0180	.0098	.0150	.0230	.0140	.0008	.0010	.0024	.0027	70.0	2.3	17.0	5.6	.1000
435	.0120	.0050	.0310	.0100	.0380	.0170	.0110	.0022	.0047	.0039	81.0	2.3	22.0	3.6	.0250
436	.0070	.0290	.0150	.0090	.0330	.0051	.0094	.0010	.0030	.0035	92.0	.3	18.0	.9	.2500
437	.0180	.0060	.0340	.0070	.0290	.0170	.0096	.0010	.0066	.0032	63.0	.3	14.0	.7	.2500
544					.0380		.0120	.0023	.0054	.0052		1.2	21.0	.7	.0250
557					.0330		.0140	.0010	.0040	.0045		.3	14.0	.6	.0250
558					.0320		.0150	.0010	.0041	.0040		.3	14.0	.6	.0250
AV	.0111	.0160	.0279	.0113	.0298	.0132	.0100	.0015	.0044	.0037	100.5	.8	19.1	1.5	.0644

Ultramafic Rocks AU ●

Ident	Sr	Ba	Cr	Zr	V	Ni	Cu	Y	Co	Sc	Zn	Pb	Ga	Sn	Ag
149-1	.0000	.0030	.2600	.0000	.0085	.0810	.0050	.0000	.0120		100.0	0.0			
185-1	.0076	.0063	.1100	.0050	.0160	.0270	.0005	.0000	.0058		49.0	0.0			
185-3	.0045	.0066	.1100	.0041	.0140	.0280	.0072	.0000	.0059		59.0	0.0			

APPENDIX 2B

Minor and trace elements

Values reported in wt. % Ba, Sr, Cr, Zr, V, Ni, Cu, Y, Co, Sc
Values reported in ppm Zn, Pb, Ga, Sn, Ag

Sunset Lake Basalt ASU ▲

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
373C	46.90	14.24	1.67	9.41	7.27	6.08	2.86	0.12	1.02	0.11	0.42	5.05	4.61	99.76	2.79
373X	42.80	15.20	1.83	10.40	8.94	7.94	1.63	0.04	0.83	0.11	0.52	5.10	4.80	100.14	2.79
374	39.68	14.35	0.54	11.30	12.07	4.50	0.90	1.03	0.82	0.15	0.56	9.71	4.30	99.91	2.79
375	50.43	17.33	1.19	7.60	6.95	6.35	2.38	0.29	1.02	0.12	0.29	1.70	3.90	99.55	2.79
376	53.11	17.46	1.19	6.10	7.07	6.44	2.43	0.19	1.01	0.11	0.23	1.41	3.70	100.45	2.79
416	47.09	17.67	1.67	4.30	10.30	5.89	2.91	0.95	0.62	0.09	0.28	4.31	3.10	99.18	2.73
417	47.25	15.84	3.70	7.50	9.22	8.35	0.31	0.22	0.72	0.07	0.28	1.83	4.80	99.91	2.73
418C	51.34	15.72	1.99	6.95	8.72	3.58	2.71	0.04	0.72	0.07	0.28	1.83	4.80	99.91	2.73
418X	50.09	14.84	1.99	7.70	9.54	4.31	2.36	0.22	0.90	0.11	0.27			92.52	2.76
419	43.23	15.18	2.36	5.80	13.20	5.42	3.19	0.32	0.92	0.11	0.30	4.10	3.60	100.13	2.76
420C	51.44	16.68	3.63	7.59	9.11	4.80	0.42	0.10	0.56	0.08	0.41	1.11	3.99	99.92	2.76
420X	52.24	15.27	3.20	8.30	8.41	5.98	0.43	0.05	0.65	0.08	0.47	0.01	4.40	99.49	2.87
421	47.84	18.41	1.89	6.90	6.34	7.94	3.64	0.07	0.73	0.07	0.46	0.89	4.00	99.18	2.87
422	50.85	15.67	2.01	9.10	7.67	6.39	2.32	0.23	0.92	0.11	0.41	0.25	3.70	99.63	2.93
423	41.89	15.87	3.18	17.00	6.12	3.71	0.05	0.15	0.60	0.09	0.84	1.88	6.60	100.18	2.93
426	54.58	13.73	1.96	11.00	3.46	5.88	1.97	0.60	1.60	0.06	0.24	0.39	4.00	100.07	2.83
427	52.81	14.06	2.30	11.70	3.12	6.48	1.78	0.50	1.63	0.69	0.26	0.18	4.70	100.21	2.83
428	51.98	15.56	1.11	7.60	7.27	7.95	2.70	1.11	0.87	0.17	0.17	0.27	2.40	99.16	2.83
429	51.50	15.03	1.43	8.70	5.54	9.52	2.09	1.04	0.80	0.15	0.17			95.97	2.83
435	51.48	14.97	1.87	7.90	8.79	4.75	2.63	0.20	1.00	0.12	0.32	2.40	3.20	99.64	2.76
436	40.04	14.58	0.87	11.90	11.68	4.74	0.86	1.03	0.86	0.15	0.57	8.20	4.30	99.78	2.79
437	47.18	17.79	2.07	6.00	10.22	6.35	3.70	0.08	0.67	0.08	0.44	2.10	2.70	99.38	2.87
544	53.40	18.60	1.20	7.10	6.17	5.64	2.90	0.79	1.21	0.10	0.22	0.20	2.30	99.83	2.79
557	49.40	14.00	1.40	10.80	10.90	7.66	2.40	0.22	1.00	0.08	0.22	0.00	1.60	99.68	2.82
558	50.90	14.00	1.30	10.00	11.10	6.59	3.10	0.19	0.86	0.10	0.20	0.40	1.50	100.24	2.82
AV	48.78	15.68	1.90	8.75	8.37	6.21	2.11	0.39	0.92	0.15	0.36	2.50	3.71		

Ultramafic Rocks AU ●

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
149-10	45.80	4.70	4.50	5.10	8.85	23.65	0.00	0.00	0.17	0.02	0.22	1.10	4.90	99.50	
185-1	49.00	11.00	2.10	8.40	9.44	14.44	1.50	0.10	0.41	0.03	0.21	0.10	3.10	100.10	
185-3	49.10	10.50	2.30	8.30	11.70	13.66	0.90	0.36	0.34	0.03	0.18	0.20	2.40	100.20	

APPENDIX 2B (cont.)

SUNSET LAKE SUBAREA

Rhyolite Ar

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
384	81.37	10.98	0.52	0.00	1.30	0.13	0.62	2.97	0.10	0.04	0.02	0.96	0.80	99.81	2.68
412	82.21	9.89	0.51	0.20	0.72	0.40	4.58	0.73	0.10	0.04	0.01	0.79	0.10	99.84	2.72
413C	77.70	10.35	0.51	0.83	1.62	0.96	1.80	2.06	0.11	0.01	0.03	2.33	1.22	99.53	2.72
413X	75.26	13.23	0.00	1.30	1.83	1.03	1.80	2.25	0.10	0.04	0.03	2.30	1.00	100.17	2.72
433	77.09	10.81	0.18	1.10	2.07	1.26	1.44	2.20	0.10	0.03	0.04	2.49	0.80	99.61	2.72
562	77.40	11.50	0.40	1.40	0.92	1.11	2.70	3.80	0.09	0.03	0.04	1.58	0.70	100.69	2.49
AV	78.51	11.13	0.28	0.81	1.41	0.82	2.16	2.34	0.10	0.03	0.04	0.60	0.77	100.69	2.49

Alice Formation - Dacite Member AAL-d

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
385	73.62	13.42	0.91	1.30	2.27	1.18	1.53	4.06	0.24	0.07	0.04	0.15	0.80	99.59	2.70
398	51.12	13.10	1.36	13.70	7.18	3.73	2.17	2.16	2.36	0.40	0.26	0.00	1.90	99.44	2.68
534	55.60	16.40	1.90	4.40	9.23	2.55	3.70	0.52	1.07	0.25	0.26	2.30	2.20	100.38	2.83
535	55.70	19.60	0.10	3.80	7.10	2.54	4.80	0.97	0.77	0.14	0.13	2.90	2.30	100.85	2.73
536	59.20	16.10	0.70	4.70	3.82	2.54	4.10	0.95	1.15	0.24	0.14	2.10	2.60	98.75	2.73
542	62.00	17.30	0.10	4.60	4.14	2.32	4.60	0.95	1.15	0.24	0.14	2.10	2.60	98.75	2.73
543	61.20	16.20	0.10	5.00	5.79	2.87	3.60	0.60	1.23	0.31	0.14	0.50	2.00	99.54	2.74
547	46.70	14.20	0.60	8.60	9.83	3.96	4.40	0.32	1.64	0.42	0.26	0.90	3.60	100.49	2.76
548	55.09	18.20	0.30	3.40	4.07	2.36	6.00	1.32	0.85	0.14	0.11	1.30	2.10	100.15	2.71
549	61.70	15.00	0.00	5.70	4.85	2.09	4.50	0.62	0.77	0.29	0.13	2.30	3.20	100.57	2.68
552	57.00	15.80	0.20	7.90	3.90	3.12	4.70	0.83	1.17	0.20	0.15	2.40	3.20	100.57	2.68
556	59.80	14.40	0.10	5.70	5.19	3.83	3.50	1.47	0.79	0.32	0.17	3.20	2.50	98.97	2.73
AV	58.57	15.82	0.53	5.73	5.61	2.63	3.97	1.21	1.08	0.26	0.15	1.94	2.42	98.97	2.72

Alice Formation - Andesite Member AAL-a

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
381	53.76	14.15	1.94	5.60	8.07	4.63	3.67	0.18	1.04	0.24	0.14	3.25	3.00	99.67	2.77
382	57.51	15.36	1.87	6.10	5.12	5.09	4.03	0.11	1.15	0.23	0.12	3.25	3.00	100.33	2.77
383	48.46	16.84	2.27	6.30	8.26	6.39	2.35	0.91	0.85	0.18	0.19	2.79	3.50	99.23	2.92
408	46.74	19.23	0.48	7.50	2.39	7.31	2.97	1.85	1.08	0.21	0.16	3.00	3.00	100.12	2.76
410	62.32	14.63	0.00	6.00	5.90	3.88	2.29	3.84	1.20	0.20	0.14	2.27	4.00	100.38	2.76
411	64.99	14.85	0.14	4.80	4.96	3.58	3.78	1.03	0.55	0.11	0.13	1.15	0.90	100.26	2.75
430	64.09	13.73	0.62	5.00	5.03	4.33	4.14	0.49	1.10	0.17	0.11	0.01	1.00	100.05	2.75
431	60.05	12.44	1.09	6.10	5.41	5.74	3.42	0.78	1.09	0.25	0.18	0.00	1.30	100.25	2.80
537	57.50	15.80	0.60	6.20	3.39	4.36	4.70	1.66	0.87	0.16	0.11	1.50	2.60	99.43	2.80
538	52.70	14.50	1.70	7.60	5.32	5.22	2.80	2.74	1.17	0.23	0.15	3.50	3.60	101.33	2.79
541	58.00	15.90	0.40	6.50	3.40	3.28	3.00	2.37	1.22	0.61	0.14	1.10	1.20	99.72	2.83
545	59.00	16.50	1.50	6.50	3.40	3.28	3.00	2.37	1.22	0.61	0.14	1.10	1.20	99.72	2.83
546	56.60	14.10	0.70	6.10	2.97	6.86	3.00	1.79	1.03	0.19	0.11	0.30	2.70	100.77	2.83
550	65.90	12.80	0.00	4.10	2.22	3.31	5.40	0.57	0.63	0.20	0.14	2.10	3.90	99.24	2.72
551	62.10	14.70	0.60	6.30	4.81	2.18	4.80	0.49	1.05	0.49	0.09	2.70	1.70	99.50	2.64
553	49.10	15.20	0.90	9.50	5.05	6.05	2.80	1.27	1.37	0.24	0.25	3.20	4.50	98.83	2.80
554	47.30	15.00	1.10	10.90	10.20	3.87	2.20	0.36	0.64	0.11	0.56	7.90	4.00	100.64	2.73
555	57.40	17.00	0.10	7.10	7.05	3.67	3.10	1.84	1.22	0.59	0.25	1.30	1.60	100.52	2.76
570	54.00	15.90	1.10	7.00	7.48	3.42	4.70	0.07	1.15	0.21	0.18	3.10	3.10	101.01	2.77
571	60.10	12.80	1.00	2.60	10.70	1.59	3.90	0.22	0.37	0.21	0.14	4.50	1.30	100.03	2.72
AV	56.80	14.93	0.91	6.61	5.48	4.52	3.55	1.17	1.00	0.24	0.17	2.17	2.45	100.03	2.77

TUMPLINE LAKE SUBAREA

Sharrise Rhyolite ASH

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
559	80.20	10.80	1.00	0.30	0.39	1.39	1.90	3.12	0.10	0.00	0.02	0.00	1.20	100.42	2.68
560	76.30	11.80	0.20	1.40	0.26	0.21	4.80	2.75	0.09	0.04	0.04	0.00	0.40	98.29	2.79
AV	75.77	12.20	0.47	1.67	0.98	1.45	3.00	3.08	0.09	0.04	0.04	0.00	0.73	98.29	2.79

Turnback Rhyolite ATB

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
378	69.43	11.01	0.00	1.80	7.07	0.33	5.40	0.22	0.99	0.57	0.18	2.90	0.40	100.30	2.67
393	76.14	12.34	0.45	1.50	2.11	1.11	3.10	2.30	0.13	0.08	0.05	0.01	0.80	100.12	2.68
394	69.23	13.15	0.73	4.90	2.79	1.86	0.47	3.69	0.61	0.19	0.18	0.16	2.10	100.06	2.68
539	77.90	12.30	0.20	1.60	0.64	1.70	1.80	3.39	0.15	0.04	0.04	0.00	1.20	100.36	2.73
561	75.90	11.80	0.90	0.90	0.68	0.50	3.50	4.17	0.14	0.03	0.02	0.20	0.40	99.14	2.65
563	76.50	11.40	0.70	0.70	0.20	0.98	1.10	6.63	0.10	0.03	0.02	0.00	0.70	99.06	2.65
564	81.50	9.60	0.30	1.30	2.26	0.62	1.40	2.27	0.13	0.03	0.04	0.10	0.70	100.25	2.69
565	72.50	12.80	1.10	1.80	0.75	0.80	4.10	3.76	0.23	0.05	0.04	0.00	0.30	98.23	2.68
572	77.50	11.50	0.10	1.60	1.56	1.86	3.10	2.10	0.14	0.04	0.04	0.00	0.70	100.24	2.67
AV	75.65	11.49	0.54	1.60	1.97	0.88	2.71	3.14	0.29	0.12	0.07	0.42	0.84	100.24	2.68

Tumpline Basalt - Andesite Member ATM-a

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
540	63.80	15.80	0.40	3.30	4.95	2.47	6.00	0.78	1.03	0.19	0.11	1.20	0.70	100.73	2.72
567	64.10	12.10	0.40	8.60	4.67	1.58	4.10	0.81	1.50	0.57	0.23	0.90	1.10	100.66	2.79
568	61.40	12.20	3.80	8.40	7.26	1.68	2.40	0.47	1.39	0.56	0.25	0.10	1.10	101.01	2.95
569	69.20	9.80	0.00	3.80	6.70	0.88	3.00	0.30	1.18	0.47	0.21	3.10	0.70	99.31	2.72
AV	64.63	12.48	1.15	6.03	5.90	1.65	3.88	0.59	1.28	0.45	0.20	1.33	0.90	99.31	2.80

Tumpline Basalt ATM

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
395	50.22	14.83	1.28	10.50	10.49	4.78	3.08	0.80	1.62	0.32	0.22	0.04	1.70	99.88	
396	54.12	15.56	1.15	9.30	9.39	4.06	2.33	0.45	1.26	0.25	0.19	0.00	1.40	99.46	
AV	52.17	15.20	1.22	9.90	9.94	4.42	2.71	0.63	1.44	0.29	0.21	0.02	1.55	99.46	

APPENDIX 2B (cont.)

CAMERON RIVER SUBAREA

Rhyolite Ar X

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
.379	67.22	14.55	1.07	3.80	2.76	2.81	3.25	2.83	0.72	0.19	0.08	0.04	1.10	100.42	2.73
.414	72.70	16.09	0.18	1.10	0.66	2.60	0.32	3.89	0.03	0.03	0.04	0.00	2.10	99.74	2.72
B1	79.10	12.60	0.00	1.30	0.90	1.30	3.50	2.50	0.02	0.02	0.01	0.00	2.10	101.23	2.65
AV	73.01	14.41	0.42	2.07	1.44	2.24	2.36	3.07	0.26	0.11	0.04	0.02	1.60		2.70

Webb Lake Andesite AW

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
.377	59.13	14.00	0.14	6.80	7.92	3.62	4.10	0.39	1.17	0.44	0.17	1.14	1.10	100.12	2.85
.415	57.49	15.12	0.59	7.60	7.98	5.33	2.07	1.21	0.86	0.18	0.15	0.34	1.70	100.62	2.69
B3	52.80	15.90	1.33	8.70	7.90	2.90	4.40	0.20	1.70		0.17			96.00	2.94
B4	52.80	16.90	0.24	6.80	4.40	3.00	6.10	0.20	0.92		0.16			95.25	2.78
B5	55.80	16.60	1.23	5.00	4.80	4.40	5.30	0.10	0.86		0.16			98.02	2.80
B10	62.90	14.00	1.40	6.20	7.10	4.40	4.00	0.70	1.10		0.13			99.60	2.82
B11	60.30	15.00	1.87	6.60	6.00	3.20	4.10	0.40	1.10		0.12			97.68	2.95
B12	57.90	14.30	1.60	6.20	7.10	4.40	4.00	0.70	1.10		0.13				2.83
AV	57.89	15.23	1.05	6.76	6.46	3.71	4.30	0.56	1.13	0.31	0.18	0.74	1.40		

Cameron River Basalt AC

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
.380	51.67	16.93	1.17	8.00	11.34	4.11	2.59	0.17	1.17	0.13	0.25	1.33	1.40	100.26	2.87
.386	51.21	14.53	1.06	10.20	10.48	7.13	2.62	0.33	0.95	0.12	0.21	0.01	1.50	100.35	2.05
.387	50.25	6.65	0.93	13.10	12.77	11.63	0.54	0.20	0.60	0.06	0.30	0.64	2.20	99.87	2.05
.388	49.34	12.16	1.56	11.50	12.79	8.26	1.02	0.16	1.19	0.12	0.24	0.00	1.60	99.94	2.04
.389	47.14	12.77	1.23	11.60	12.00	9.62	1.20	0.33	1.38	0.11	0.25	0.01	2.10	99.74	2.04
.390	54.55	13.81	1.78	8.00	10.96	3.48	3.48	0.12	1.58	0.20	0.22	0.55	0.90	99.63	2.94
.391	53.04	14.91	1.52	11.00	9.28	4.06	2.95	0.25	1.81	0.22	0.29	0.00	1.20	100.53	2.92
.392	50.17	15.21	1.44	8.00	13.20	5.90	1.33	0.13	0.84	0.10	0.22	0.32	1.50	100.36	2.98
.400	50.38	17.17	1.02	11.20	12.17	3.44	0.96	0.95	1.07	0.10	0.31	2.31	1.10	99.85	3.04
.401	49.27	16.15	0.82	11.20	10.64	5.67	3.60	0.31	1.03	0.12	0.25	0.02	1.40	99.77	3.05
.402C	49.56	15.05	2.50	10.48	10.33	6.42	2.31	0.16	1.05	0.11	0.21	0.02	1.41	100.20	3.05
.402X	50.58	15.02	1.57	11.00	10.45	5.88	2.53	0.11	1.12	0.11	0.22	0.01	1.50	99.97	3.05
.403	51.38	14.97	0.01	11.70	10.45	5.94	2.53	0.11	1.12	0.11	0.22	0.06	1.40	101.72	2.96
.404	55.08	14.57	0.00	11.35	9.90	4.03	1.71	0.21	1.22	0.13	0.37	1.90	1.25	100.73	2.96
.405	51.54	14.11	1.25	11.40	11.70	4.35	1.47	0.19	1.12	0.13	0.32	1.90	1.25	99.34	2.05
.406	50.00	13.97	1.23	10.80	11.95	6.26	0.94	0.13	0.95	0.10	0.29	0.92	1.60	99.32	2.05
.407	45.62	14.87	1.02	10.80	11.95	5.47	1.56	0.29	0.94	0.11	0.28	2.90	1.60	100.47	3.04
.424	54.51	14.93	1.41	8.60	11.95	3.80	1.36	0.11	1.08	0.09	0.31	1.12	1.20	99.31	3.04
.425	49.87	16.62	0.91	10.40	11.48	4.24	1.78	0.13	1.19	0.15	0.40	0.94	1.20	99.72	2.83
.434X	51.46	14.55	1.86	10.50	11.25	5.54	2.01	0.11	1.10	0.11	0.20			93.64	2.86
.B6	48.00	14.80	1.77	9.30	9.50	5.90	2.80	0.10	1.10		0.21			96.76	2.80
.B7	53.80	14.60	1.93	7.80	7.80	2.90	2.90	0.60	1.10		0.36			97.16	2.95
.B8	48.20	14.60	3.20	11.60	10.10	4.90	2.50	0.20	1.10		0.18				
.B14	50.80	13.50	1.10	10.70	9.70	7.50	2.50	0.40	0.78		0.26	0.78	1.43		
AV	50.73	14.44	1.35	10.38	11.06	5.69	2.04	0.24	1.11	0.12	0.18				

Mafic Dykes Aa X

Ident	SiO2	Al2O3	Fe2O3	FeO	CaO	MgO	Na2O	K2O	TiO2	P2O5	MnO	CO2	H2O	Total	S.G.
.399	47.24	14.36	0.24	11.60	10.93	10.03	1.39	0.20	0.62	0.08	0.21	0.05	2.70	99.65	3.02
.432	51.05	13.76	1.08	12.50	8.03	4.27	2.60	1.15	3.06	0.40	0.21	0.00	1.00	99.11	
.566	48.50	16.00	1.50	11.40	7.52	6.38	2.90	0.78	2.26	0.32	0.21	0.00	2.10	99.87	2.99
.B2	49.10	15.20	1.60	10.00	9.40	6.60	2.40	0.30	0.88		0.16			95.64	2.99
.B9	52.70	13.40	2.80	10.90	7.50	4.50	2.30	0.10	1.20		0.17			95.57	2.85
.B13	48.60	15.00	3.80	10.00	9.40	7.00	2.40	0.20	0.89		0.17			97.46	2.92
AV	49.53	14.62	1.84	11.07	8.80	6.46	2.33	0.46	1.49	0.27	0.19	0.02	1.93		2.95

APPENDIX 2C

Normative compositions (Barth-Niggli cation %)

RU zero except for following samples
Rhyolite (10) — 384 — 0.06 (Sunset Lake subarea)

$$D.I. \text{ Differentiation index} = QTZ + AB + OR + NE + KP + LC$$

HM zero except for following samples
Tumpline Rhyolite (11) — 559 — 0.35
Rhyolite (10) — 384 — 0.38 (Sunset Lake subarea)

$$PLAG \text{ Normative plagioclase composition} = \frac{100 \text{ An}}{\text{An} + \text{Ab} + 5/3\text{Ne}}$$

C.I. Normative colour index = OL+OPX + CPX + MT + HM

CAMERON RIVER SUBAREA																		
Rhyolite Ar X																		
Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C.I.	D.I.
379	24.21	1.85	17.01	29.65	12.40	0.00	0.00	7.88	4.33	0.00	0.00	1.14	1.02	0.40	0.10	29.48	14.37	70.87
414	48.03	11.70	24.12	3.01	3.23	0.00	0.00	7.53	1.68	0.00	0.00	0.20	0.04	0.07	0.00	51.73	9.44	75.56
B1	40.96	2.73	14.79	31.43	4.47	0.00	0.00	3.59	2.00	0.00	0.00	0.00	0.03	0.00	0.00	12.44	5.62	87.19
AV	37.87	5.43	18.64	21.37	6.70	0.00	0.00	6.33	2.67	0.00	0.00	0.45	0.36	0.16	0.03	31.22	9.81	77.87
Webb Lake Andesite AW																		
Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C.I.	D.I.
377	12.33	0.00	2.33	37.25	18.87	4.48	4.07	7.87	7.15	0.00	0.00	0.15	1.65	0.93	2.92	33.63	25.36	51.91
415	12.59	0.00	7.32	19.01	29.04	3.78	2.67	13.16	9.30	0.00	0.00	0.63	1.23	0.39	0.88	60.44	30.77	38.92
83	2.95	0.00	1.24	41.48	24.21	5.89	7.68	5.47	7.13	0.00	0.00	1.46	2.49	0.00	0.00	36.85	30.11	45.68
84	2.30	0.00	1.22	56.49	18.72	1.43	1.61	7.83	8.83	0.00	0.00	0.26	1.32	0.00	0.00	24.89	21.27	60.01
85	4.09	0.00	0.61	49.31	21.99	4.23	0.82	11.92	7.35	0.00	0.00	1.33	1.24	0.00	0.00	30.84	24.00	54.01
B10	18.20	0.00	2.44	39.31	18.24	4.65	3.64	5.63	4.40	0.00	0.00	1.51	1.58	0.00	0.00	31.46	21.40	60.36
B11	11.38	0.00	7.78	37.27	18.92	4.83	4.15	6.53	5.61	0.00	0.00	1.98	1.55	0.00	0.00	33.67	24.65	56.43
B12	9.45	0.00	4.25	36.90	19.52	8.48	4.86	8.24	4.72	0.00	0.00	1.72	1.86	0.00	0.00	34.60	29.88	50.60
AV	9.16	0.00	3.40	39.68	21.19	4.36	3.69	8.33	6.81	0.00	0.00	1.13	1.61	0.17	0.48	35.80	25.93	52.24
Cameron River Basalt AC																		
Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C.I.	D.I.
380	6.37	0.00	1.03	23.75	34.80	5.53	5.04	8.82	8.04	0.00	0.00	1.25	1.66	0.28	3.44	59.44	30.34	31.15
386	0.00	0.00	1.99	23.95	27.41	11.51	8.23	13.47	9.63	0.00	0.00	1.13	1.35	0.26	0.03	53.37	46.37	25.94
387	2.52	0.00	1.23	23.95	27.41	11.51	8.23	13.47	9.63	0.00	0.00	1.13	1.35	0.26	0.03	53.37	46.37	25.94
388	2.44	0.00	0.98	9.52	29.25	17.13	11.60	15.14	10.26	0.00	0.00	1.70	1.72	0.26	0.00	75.44	37.55	12.94
389	0.00	0.00	2.02	11.17	29.54	15.83	9.29	13.10	7.69	4.89	2.87	1.33	1.99	0.24	0.03	72.56	37.00	13.20
390	7.85	0.00	0.73	32.12	22.33	11.66	11.28	4.05	3.92	0.00	0.00	1.94	2.26	0.43	1.43	41.00	35.09	40.72
391	5.69	0.00	1.52	27.29	27.52	6.98	8.17	8.26	10.25	0.00	0.00	1.64	2.60	0.48	0.00	50.21	37.50	34.50
392	6.85	0.00	0.80	12.43	36.60	12.06	7.82	10.93	7.09	0.00	0.00	1.97	1.22	0.22	0.05	47.96	34.95	34.58
400	0.00	0.00	1.86	32.73	30.16	7.44	6.24	4.40	3.59	5.80	4.86	1.08	1.45	0.25	0.05	47.96	34.95	34.58
401	8.36	0.00	5.84	8.96	38.43	2.55	4.24	8.60	14.28	0.00	0.00	0.89	1.55	0.22	6.07	81.09	32.12	23.16
402C	1.13	0.00	0.98	21.43	31.24	9.44	7.25	13.60	10.44	0.00	0.00	2.70	1.51	0.24	0.03	60.64	44.50	23.56
402X	2.33	0.00	0.67	20.56	31.68	9.69	8.80	11.90	10.80	0.00	0.00	1.69	1.61	0.24	0.16	56.22	44.65	24.99
403	0.98	0.00	0.67	23.34	29.97	8.75	8.99	12.47	12.82	0.00	0.00	0.01	1.60	0.24	0.16	56.22	44.65	24.99
404	13.99	0.00	1.27	15.66	32.09	1.58	2.33	10.56	15.62	0.00	0.00	0.00	1.73	0.28	4.90	67.21	31.82	30.91
405	10.29	0.00	1.16	13.63	31.37	4.95	6.48	9.92	13.00	0.00	0.00	1.35	1.39	0.24	2.44	79.89	45.24	17.02
406	7.39	0.00	0.81	8.83	35.08	8.49	7.67	13.84	12.51	0.00	0.00	1.35	1.39	0.24	2.44	79.89	45.24	17.02
407	1.25	0.00	1.78	14.49	33.86	7.02	7.05	12.11	12.15	0.00	0.00	1.10	1.35	0.24	2.44	79.89	45.24	17.02
424	15.19	0.00	0.67	12.64	35.53	6.90	7.44	7.41	8.00	0.00	0.00	1.53	1.56	0.20	2.47	69.83	35.37	23.33
425	9.89	0.00	0.80	16.64	38.50	4.87	6.01	9.75	12.04	0.00	0.00	0.99	1.73	0.33	0.00	62.72	43.98	24.34
434X	4.98	0.00	0.67	18.69	31.44	10.78	9.72	10.45	9.43	0.00	0.00	2.01	1.59	0.24	0.00	52.27	42.54	27.76
86	0.00	0.00	0.64	27.12	29.70	9.76	7.17	12.60	9.25	0.00	0.05	2.00	1.55	0.24	0.00	52.27	42.54	27.76
87	11.51	0.00	3.87	28.38	27.31	5.43	6.47	6.01	7.16	0.00	0.00	2.20	1.67	0.00	0.00	49.04	28.94	43.75
88	0.07	0.00	1.25	23.79	29.71	8.57	10.15	10.25	11.90	0.00	0.00	2.88	1.62	0.00	0.00	55.54	45.18	25.11
B14	0.00	0.00	2.45	23.27	25.33	11.39	8.26	15.27	11.07	0.37	0.27	1.19	1.13	0.00	0.00	52.13	48.95	25.72
AV	4.80	0.00	1.49	18.98	30.65	9.20	7.89	11.03	10.18	0.49	0.35	1.44	1.60	0.22	1.70	62.93	42.18	25.26
Mafic Dykes Aa X																		
Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C.I.	D.I.
399	0.00	0.00	1.22	12.90	33.46	10.71	6.68	12.89	8.04	7.78	4.85	0.26	0.89	0.17	0.13	72.17	52.12	14.12
432	4.22	0.00	7.14	24.51	23.61	5.52	6.87	9.62	11.98	0.00	0.00	1.19	4.48	0.88	0.00	49.06	39.64	35.97
566	0.00	0.00	4.77	26.94	29.33	3.19	2.50	12.25	9.59	3.28	2.57	1.62	3.26	0.69	0.00	52.12	38.27	31.71
82	0.40	0.00	1.87	22.74	31.47	8.18	6.02	15.14	11.13	0.00	0.00	1.77	1.29	0.00	0.00	58.06	43.52	25.01
89	11.17	0.00	0.64	22.27	27.99	4.62	5.10	11.09	12.27	0.00	0.00	3.05	1.80	0.00	0.00	55.69	37.93	34.08
B13	0.00	0.00	1.23	22.43	30.79	8.05	6.17	13.90	10.64	1.64	1.26	2.60	1.29	0.00	0.00	57.85	45.55	23.66
AV	2.63	0.00	2.81	21.97	29.44	6.71	5.56	12.48	10.61	2.12	1.45	1.75	2.17	0.29	0.02	57.49	42.84	27.41

APPENDIX 2C (cont.)

TUMPLINE LAKE SUBAREA

Sharnie Rhyolite ASH +

Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C I	D I
559	52.11	4.06	19.12	17.68	2.01	0.00	0.00	3.98	0.00	0.00	0.00	0.56	0.14	0.00	0.00	10.19	5.03	88.91
560	34.29	0.61	16.71	44.28	1.06	0.00	0.00	4.12	2.02	0.00	0.00	0.22	0.13	0.09	0.00	2.33	2.96	35.28
AV	39.25	2.52	18.76	27.66	4.71	0.00	0.00					0.50	0.25	0.09	0.00	14.97	7.01	85.67

Turnback Rhyolite ATB Δ

Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C I	D I
378	29.47	0.00	1.30	48.44	5.15	1.82	3.38	0.00	0.00	0.00	0.00	0.00	1.38	1.19	7.33	9.62	6.58	79.21
393	40.19	1.26	13.94	28.52	10.13	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.19	0.17	0.03	26.20	5.76	62.65
394	40.36	4.79	23.16	4.48	12.30	0.00	0.00	5.45	6.91	0.00	0.00	0.81	0.90	0.42	0.03	73.31	14.07	67.99
559	46.76	5.19	20.71	16.70	3.01	0.00	0.00	4.85	2.27	0.00	0.00	0.22	0.22	0.06	0.00	16.28	7.55	84.17
561	35.79	0.92	25.30	32.24	1.96	0.00	0.00	1.42	0.62	0.00	0.00	0.97	0.20	0.06	0.00	5.74	3.20	93.33
563	41.21	2.41	40.93	10.31	0.83	0.00	0.00	2.82	0.51	0.00	0.00	0.76	0.14	0.07	0.00	7.46	4.24	92.45
564	56.43	1.18	14.02	13.13	10.85	0.00	0.00	1.79	1.76	0.00	0.00	0.33	0.19	0.07	0.00	45.24	4.07	83.58
565	29.20	0.83	22.89	37.88	3.49	0.00	0.00	2.27	1.82	0.00	0.00	1.18	0.33	0.11	0.00	8.44	5.60	89.97
572	41.78	1.54	12.68	28.41	7.63	0.00	0.00	5.24	2.32	0.00	0.00	0.11	0.20	0.09	0.00	21.18	7.87	82.81
AV	41.23	1.90	19.31	24.88	5.54	0.23	0.42	2.40	1.72	0.00	0.00	0.52	0.41	0.26	1.07	22.51	5.74	85.43

Tumpline Basalt - Andesite Member ATM-a ●

Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C I	D I
540	12.67	0.00	4.54	52.99	13.65	0.95	0.50	6.23	3.26	0.00	0.00	0.41	1.41	0.39	2.99	20.49	12.77	70.20
567	21.93	0.00	4.92	37.77	12.54	0.35	0.00	5.23	1.16	0.00	0.00	0.43	2.14	1.22	2.33	24.93	19.28	64.61
568	25.17	0.00	2.91	22.57	22.14	2.68	6.20	3.52	8.12	0.00	0.00	0.00	0.00	0.00	0.27	49.52	25.71	50.66
569	25.09	0.76	1.84	27.89	10.94	0.00	0.00	2.43	4.74	0.00	0.00	0.00	1.70	1.23	0.12	28.16	15.10	63.30
AV	25.09	0.19	3.55	35.30	14.82	1.00	1.90	2.43	6.82	0.00	0.00	1.00	1.82	0.37	3.43	30.78	16.66	63.94

Tumpline Basalt ATM □

Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C I	D I
395	0.00	0.00	4.89	28.56	25.08	10.40	10.61	6.30	6.42	1.60	1.63	1.38	2.33	0.69	0.10	46.76	40.67	33.45
396	9.56	0.00	2.77	31.79	31.96	5.71	6.19	9.56	8.82	0.00	0.00	1.25	1.83	0.55	0.00	53.46	33.37	33.45
AV	4.78	0.00	3.83	25.18	28.52	8.06	8.40	7.56	7.99	0.80	0.82	1.32	2.08	0.62	0.05	53.11	37.02	33.79

SUNSET LAKE SUBAREA

Rhyolite Ar ■

Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C I	D I
384	64.39	7.69	18.46	5.85	0.13	0.00	0.00	0.38	0.00	0.00	0.00	0.00	0.03	0.09	2.55	2.09	0.79	88.70
412	49.28	1.73	4.38	41.70	0.00	0.00	0.00	0.51	0.06	0.00	0.00	0.07	0.14	0.09	0.00	0.00	0.79	88.70
413C	55.90	5.87	12.70	16.81	0.00	0.00	0.00	0.51	0.19	0.00	0.00	0.56	0.16	0.02	0.00	0.00	0.79	88.70
413X	52.19	8.79	13.67	16.81	0.00	0.00	0.00	1.52	1.03	0.00	0.00	0.00	0.14	0.09	0.00	0.00	2.69	82.46
433	56.49	6.86	13.49	13.41	0.00	0.00	0.00	1.52	1.03	0.00	0.00	0.00	0.14	0.09	0.00	0.00	2.69	82.46
562	41.70	3.03	22.87	24.67	0.59	0.00	0.00	3.12	1.86	0.00	0.00	0.20	0.14	0.07	0.00	0.00	1.15	83.39
AV	53.49	5.66	14.26	19.85	0.12	0.00	0.00	1.35	0.67	0.00	0.00	0.21	0.13	0.07	0.00	0.00	2.41	87.60

Alice Formation - Dacite Member AAL-d ◆

Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C I	D I
385	40.90	3.28	24.92	14.26	10.23	0.00	0.00	3.38	1.15	0.00	0.00	0.99	0.35	0.15	0.39	41.78	5.87	80.08
398	3.85	0.00	13.58	20.71	20.86	4.23	7.15	8.83	14.91	0.00	0.00	1.51	3.49	0.89	0.00	50.18	40.13	38.13
534	12.99	0.00	3.13	33.85	27.12	1.51	0.94	6.42	4.02	0.00	0.00	2.02	1.52	0.53	0.00	44.48	16.44	43.98
535	9.85	5.20	5.70	42.78	15.86	0.00	0.00	6.96	4.91	0.00	0.00	0.10	1.07	0.29	0.00	7.28	27.05	13.04
536	21.68	7.71	5.81	38.09	4.25	0.00	0.00	8.43	5.60	0.00	0.00	0.00	0.14	0.09	0.00	10.02	16.44	58.32
542	17.48	3.02	4.43	42.34	17.48	0.00	0.00	6.57	6.10	0.00	0.00	0.11	1.23	0.72	0.00	29.22	14.00	64.25
543	19.26	1.13	3.66	33.31	24.26	0.00	0.00	11.10	11.20	0.00	0.00	0.11	1.77	0.67	0.00	42.14	16.41	56.23
547	4.01	3.79	1.92	40.12	8.86	0.00	0.00	6.53	4.06	0.00	0.00	0.32	1.19	0.29	0.00	15.15	18.09	25.26
548	8.14	3.23	7.83	54.02	11.09	0.00	0.00	5.87	8.11	0.00	0.00	0.00	0.62	0.29	0.00	17.03	12.09	70.00
549	21.17	4.09	3.73	41.13	7.76	0.00	0.00	5.87	8.11	0.00	0.00	0.00	0.62	0.29	0.00	17.03	12.09	70.00
552	14.00	6.82	5.01	43.05	2.93	0.00	0.00	8.79	10.92	0.00	0.00	0.21	1.66	0.43	0.00	6.19	15.88	15.07
556	24.37	6.57	9.03	32.64	3.56	0.00	0.00	5.25	8.23	0.00	0.00	0.11	1.14	0.70	0.00	9.83	14.73	66.04
AV	16.48	3.78	7.40	36.36	12.85	0.48	0.67	7.19	7.13	0.00	0.00	0.57	1.54	0.56	4.99	26.01	17.59	60.23

Alice Formation - Andesite Member AAL-a +

Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fe	Fa	Mt	Il	Ap	Cc	Plag	C I	D I
381	13.14	1.52	1.10	33.88	18.43	0.00	0.00	13.14	6.27	0.00	0.00	2.08	1.49	0.52	8.45	35.23	22.98	48.11
382	15.33	3.79	0.66	36.71	14.37	0.00	0.00	14.26	6.83	0.00	0.00	2.98	1.63	0.49	13.95	28.13	24.70	52.70
383	7.94	4.38	5.58	21.87	22.98	0.00	0.00	18.29	7.56	0.00	0.00	2.46	1.71	0.39	13.95	28.13	24.70	52.70
409	15.29	9.08	11.11	27.08	0.00	0.00	0.00	18.29	7.56	0.00	0.00	0.51	1.53	0.45	7.71	0.00	33.28	39.40
410	0.74	12.57	23.14	20.95	0.00	0.00	0.00	21.72	9.04	0.00	0.00	0.81	1.70	0.43	5.85	0.00	36.33	44.83
411	16.14	0.00	6.14	34.18	20.06	0.58	0.53	9.61	8.84	0.00	0.00	0.00	0.77	0.23	0.00	31.93	20.31	56.85
430	19.05	0.00	2.94	33.99	20.56	1.73	1.17	9.14	6.16	0.00	0.00	0.15	0.86	0.21	0.00	31.93	20.31	56.85
431	18.44	0.00	2.94	33.99	20.56	1.73	1.17	9.14	6.16	0.00	0.00	0.15	0.86	0.21	0.00	31.93	20.31	56.85
537	9.61	4.41	10.00	42.96	6.40	0.00	0.00	10.56	5.31	0.00	0.00	0.66	1.55	0.57	0.00	31.93	20.31	56.85
538	11.24	6.58	16.52	25.63	2.82	0.00	0.00	14.37	8.30	0.00	0.00	1.17	1.56	0.54	0.00	31.93	20.31	56.85
541	14.33	2.90	14.98	27.64	16.26	0.00	0.00	14.69	9.35	0.00	0.00	0.64	1.23	0.34	0.00	31.93	20.31	56.85
545	15.94	2.68	10.91	27.64	16.26	0.00	0.00	10.56	5.31	0.00	0.00	0.43	1.74	1.31	0.00	31.93	20.31	56.85
546	16.18	7.81	4.89	35.24	0.16	0.00	0.00	19.56	8.09	0.00	0.00	1.62	1.48	0.41	0.00	31.93	20.31	56.85
550	24.03	3.63	3.39	48.82	0.00	0.00	0.00	7.56	4.65	0.00	0.00	0.76	1.40	0.43	0.00	31.93	20.31	56.85
551	16.23	0.00	2.97	44.18	17.55	0.35	0.47	5.99	8.02	0.00	0.00	0.64	1.50	1.05	0.00	31.93	20.31	56.85
553	10.43	9.19	7.88	26.37	2.41	0.00	0.00	11.08	16.70	0.00	0.00	0.43	1.73	1.26	0.00	31.93	20.31	56.85
554	17.95	8.49	2.21	20.49	0.00	0.00	0.00	7.56	4.65	0.00	0.00	0.99	2.00	0.53	0.00	31.93	20.31	56.85
555	14.65	3.71	11.10	28.38	18.29	0.00	0.00	11.08	16.70	0.00	0.00	0.19	0.93	0.24	0.00	31.93	20.31	56.85
570	9.00	1.97	0.42	42.70	16.33	0.00	0.00	7.51	9.60	0.00	0.00	0.43	1.73	1.26	0.00	31.93	20.31	56.85
571	22.19	0.00	1.00	32.54	16.93	3.49	1.75	2.69	1.35	0.00	0.00	0.16	1.36	0.44	11.49	32.39	21.60	52.12
41V	14.42	3.94	7.06	32.54	12.91	0.61	0.34	12.18	8.36	0.00	0.00	0.98	1.42	0.52	5.63	23.87	21.88	54.02

APPENDIX 2C (cont.)

Sunset Lake Basalt ASU ▲

Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fo	Fa	Mt	Il	Ap	Cc	Plag	C.I.	D.I.
373C	12.02	9.25	0.74	26.79	3.57	0.00	0.00	17.51	13.20	0.00	0.00	1.82	1.48	0.24	13.33	11.76	34.02	39.55
373X	8.29	9.45	0.25	15.26	11.88	0.00	0.00	22.86	15.12	0.00	0.00	1.99	1.21	0.24	13.45	43.77	41.18	23.80
374	14.69	13.37	6.35	8.42	0.00	0.00	0.00	12.50	16.98	0.00	0.00	0.59	1.19	0.33	25.59	0.00	31.26	29.46
375	10.36	8.35	1.79	22.30	23.96	0.00	0.00	18.30	10.42	0.00	0.00	1.30	1.48	0.26	4.49	51.79	31.50	34.45
376	12.95	4.43	1.16	22.48	26.22	0.00	0.00	18.32	7.81	0.00	0.00	1.28	1.45	0.24	3.67	53.84	28.85	36.59
416	4.67	3.70	5.72	26.59	23.68	0.00	0.00	16.55	5.16	0.00	0.00	1.78	0.88	0.19	11.09	47.11	24.36	36.98
417	11.90	3.41	0.25	2.97	35.95	0.00	0.00	24.58	12.31	0.00	0.00	2.47	1.07	0.16	4.93	92.38	40.43	15.12
418C	9.24	0.00	1.43	26.70	33.01	5.86	5.08	7.91	6.86	0.00	0.00	2.28	1.38	0.25	0.00	55.29	29.37	37.82
418X	14.59	3.66	1.17	22.06	21.41	0.00	0.00	12.39	10.05	0.00	0.00	1.7	1.42	0.28	10.80	49.25	26.02	37.82
419	0.40	0.00	0.93	29.20	26.68	0.24	0.00	11.15	14.25	0.00	0.00	0.52	1.31	0.24	15.58	47.74	25.98	31.53
420C	18.46	2.44	0.64	4.06	40.51	0.00	0.00	14.25	13.28	0.00	0.00	0.32	0.84	0.18	3.02	90.90	30.69	23.15
420X	16.48	0.00	0.32	4.20	43.03	0.76	0.58	17.56	13.44	0.00	0.00	0.44	0.98	0.18	0.03	91.12	35.77	21.00
421	0.00	0.43	0.43	33.69	26.16	0.00	0.00	19.11	7.92	2.61	1.08	0.44	1.05	0.15	2.32	43.71	33.81	34.12
422	5.74	0.00	1.43	21.95	33.38	0.05	1.40	17.57	12.01	0.00	0.00	0.22	1.35	0.24	0.67	60.33	36.59	29.13
423	11.06	11.08	0.99	0.50	20.02	0.00	0.00	17.64	29.25	0.00	0.00	0.69	1.25	0.21	5.32	97.55	50.83	12.55
426	19.92	7.04	3.79	18.91	11.10	0.00	0.00	17.35	14.78	0.00	0.00	0.19	2.38	1.56	0.49	38.07	40.08	39.07
427	18.68	8.22	3.18	17.21	10.58	0.00	0.00	19.26	15.79	0.00	0.00	0.59	2.44	1.56	0.49	38.07	40.08	39.07
428	1.64	0.00	6.74	24.90	27.80	0.49	1.60	20.80	9.53	0.00	0.00	1.19	1.24	0.37	0.70	52.75	37.85	33.28
429	3.31	0.85	6.39	19.49	27.53	0.00	0.00	27.30	12.08	0.00	0.00	1.55	1.16	0.33	0.00	58.55	42.09	29.19
435	10.62	0.34	1.24	24.66	28.88	0.00	0.00	13.69	10.49	0.00	0.00	2.04	1.46	0.26	6.34	53.94	27.68	36.51
436	12.01	11.66	6.40	8.11	5.38	0.00	0.00	13.74	18.40	0.00	0.00	0.95	1.25	0.33	21.77	39.90	34.35	26.51
437	0.00	0.00	0.48	33.76	32.23	0.95	1.28	8.37	3.64	5.98	2.59	0.17	5.40	48.84	27.96	34.24	34.24	34.24
544	7.58	2.75	4.79	26.71	29.43	0.00	0.00	15.97	9.05	0.00	0.00	0.29	1.73	0.22	0.52	52.43	28.03	39.04
557	0.00	0.00	1.34	22.15	27.54	12.99	9.03	10.51	7.31	3.55	2.47	1.51	1.43	0.17	0.00	55.42	48.80	23.49
558	0.00	0.00	1.14	28.31	24.14	13.07	9.84	8.94	6.74	2.27	1.71	1.38	1.22	0.21	1.03	46.02	45.17	29.45
AV	8.98	4.02	2.40	19.66	23.76	1.66	1.16	16.33	11.53	0.58	0.31	1.87	1.34	0.34	6.06	52.78	34.78	31.04

Ultramafic Rocks AU ●

Ident	Qtz	Cor	Or	Ab	An	Di	He	En	Fs	Fo	Fa	Mt	Il	Ap	Cc	Plag	C.I.	D.I.
149-10	0.00	0.00	0.00	0.00	12.68	16.40	2.57	48.42	7.58	5.90	0.92	1.73	0.23	0.04	2.75	100.00	83.76	0.00
185-1	0.00	0.00	0.60	13.71	23.41	14.54	4.22	25.21	7.32	6.09	1.77	2.03	0.58	0.06	0.25	63.06	61.75	14.32
185-3	0.00	0.00	2.17	8.22	23.95	20.47	6.41	23.65	7.41	3.35	1.05	1.96	0.48	0.06	0.51	74.46	64.78	10.38

APPENDIX 2D

SiO₂ [variation diagrams]

Symbols refer to formations in Appendices 2A, 2B, 2C, and Figures 41, 42, 49
 Symbols in parentheses refer to Appendix 2D only

Cameron River subarea

×	Rhyolite	AR
◇ (◇)	Webb Lake Andesite	AW
○ (①)	Cameron River Basalt	AC
× (⊗)	Mafic Dykes	Aa

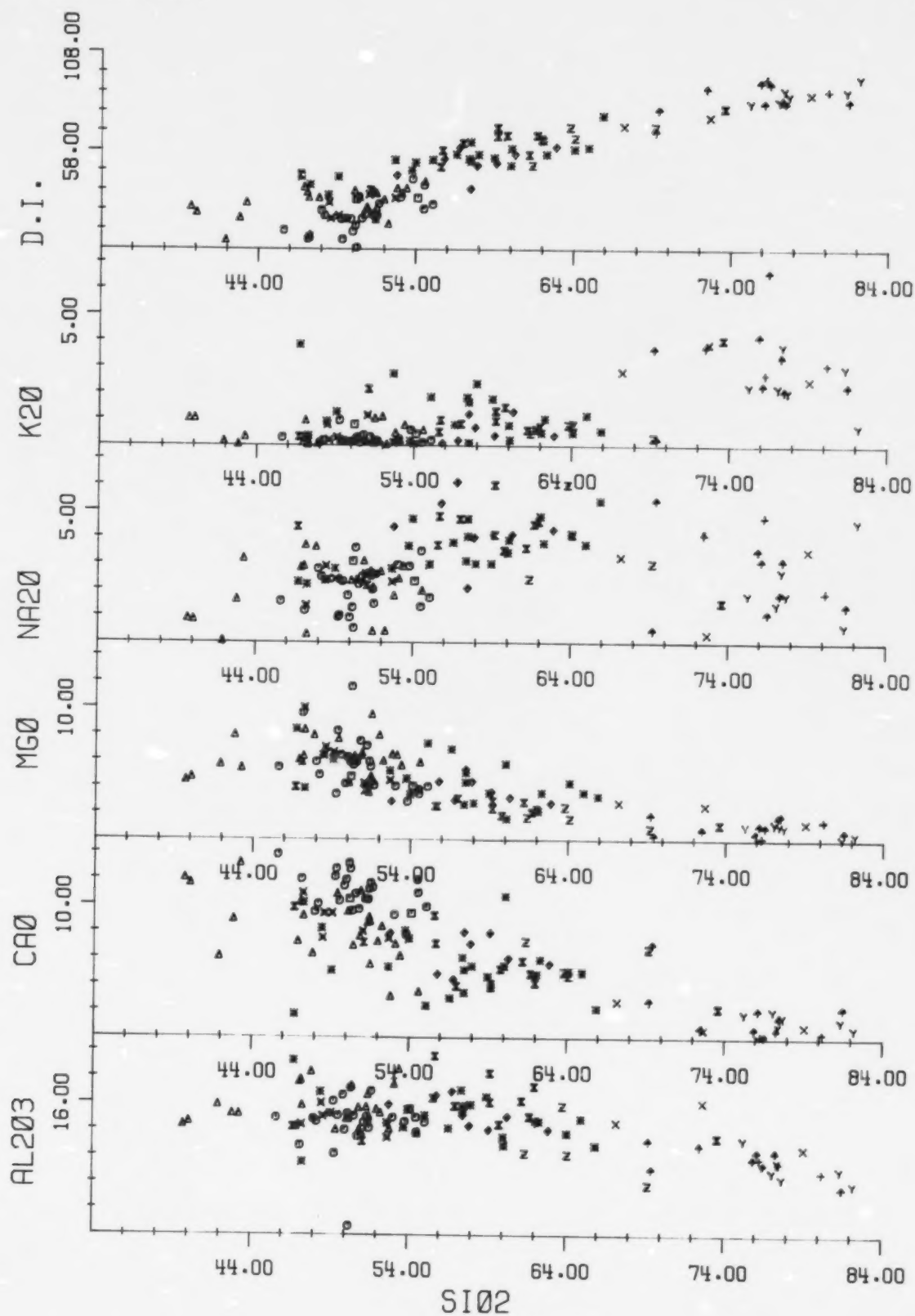
Sunset Lake subarea

■ (Y)	Rhyolite	Ar
◆ (⊗)	Alice Formation - Dacite Member	AAL-d
⊕ (⊗)	Alice Formation - Andesite Member	AAL-a
▲ (△)	Sunset Lake Basalt	ASU
●	Ultramafic Rocks	AU

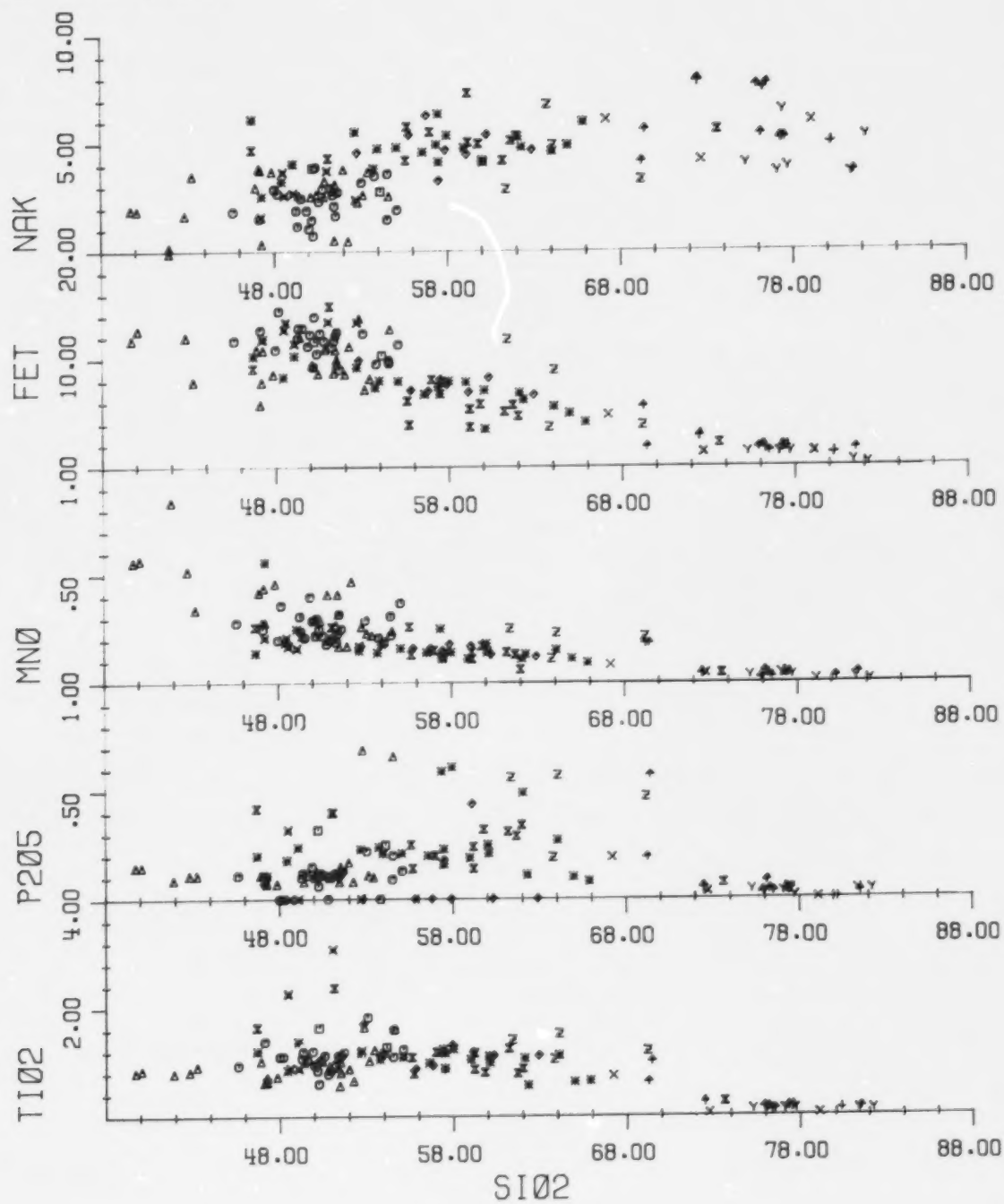
Tumpline Lake subarea

+	Sharrie Rhyolite	ASH
△ (▲)	Turnback Rhyolite	ATB
● (Z)	Tumpline Basalt - Andesite Member	ATM-a
□ (□)	Tumpline Basalt	ATM

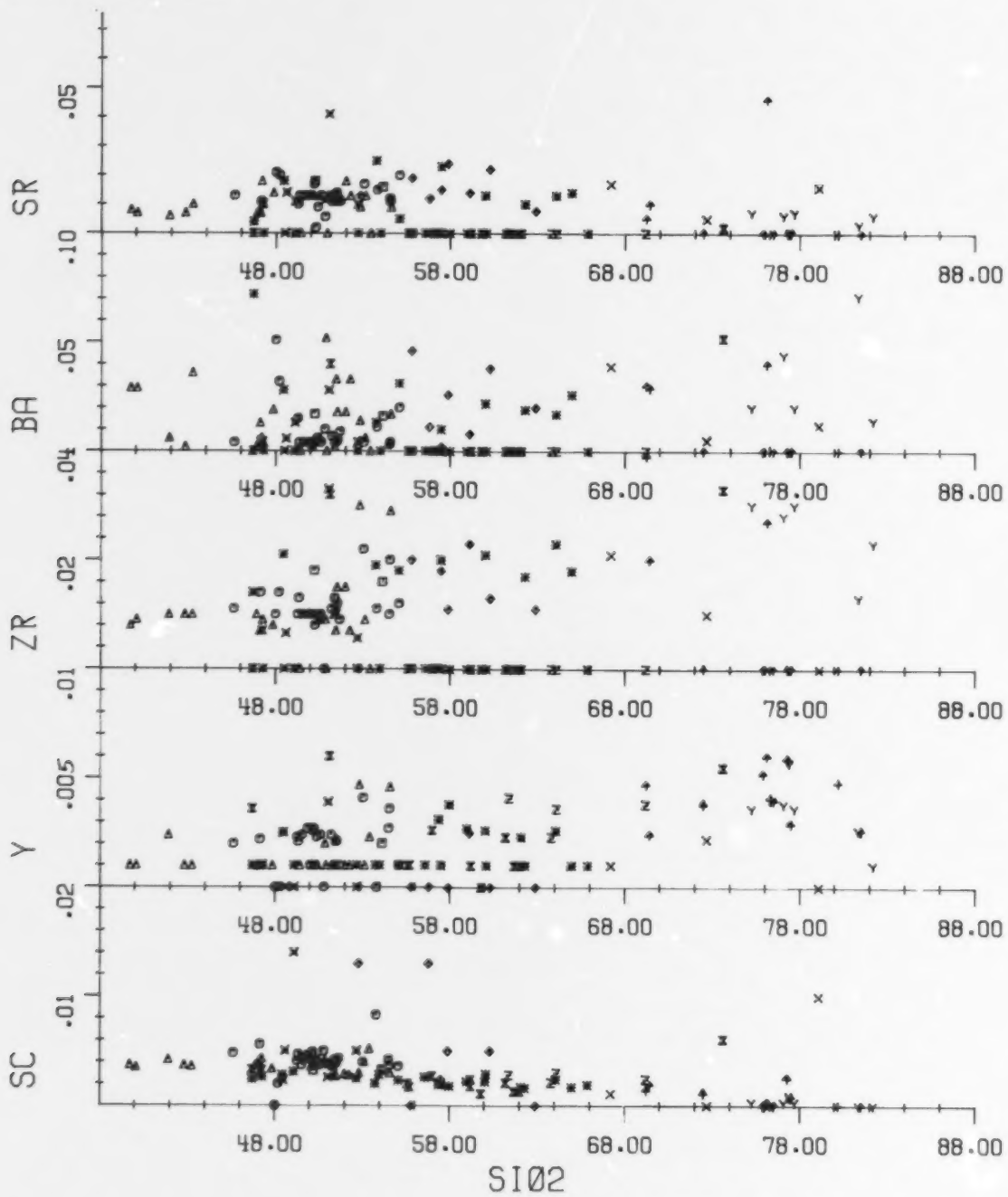
APPENDIX 2D (cont.)



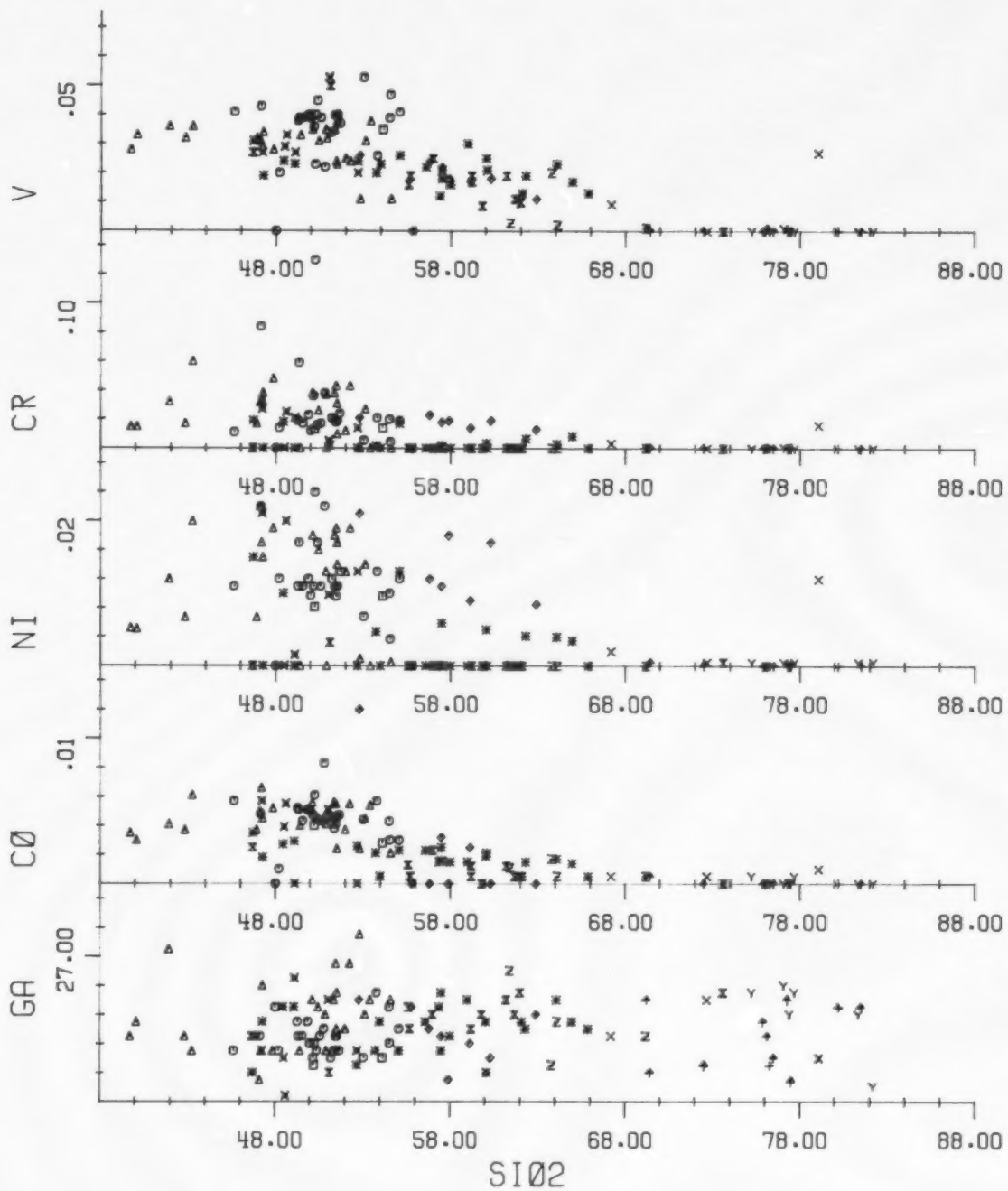
APPENDIX 2D (cont.)



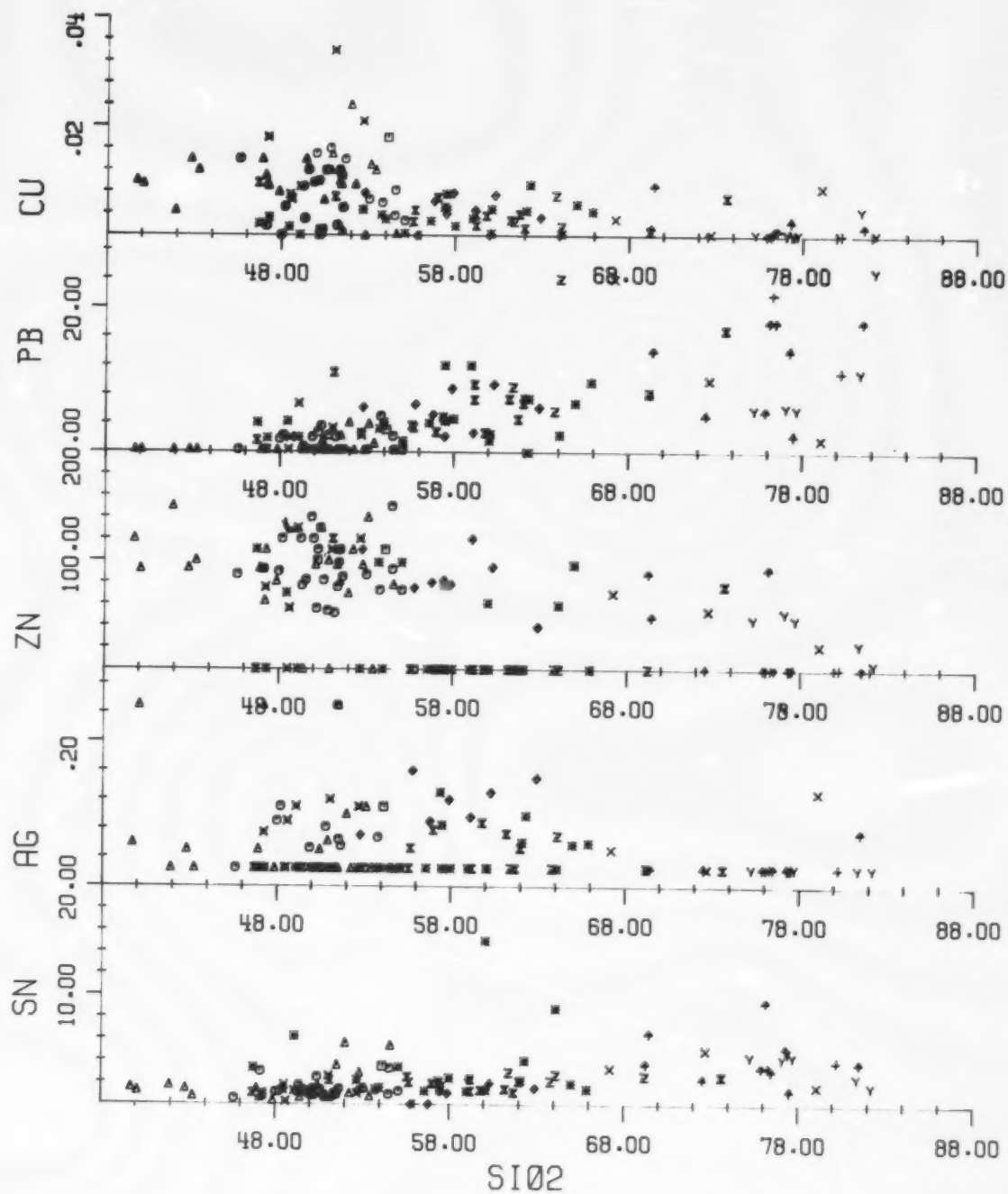
APPENDIX 2D (cont.)



APPENDIX 2D (cont.)



APPENDIX 2D (cont.)



**MAPS
NOT
FILMED**

**CARTES
OMISES**